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Edited by F. E. BEDDARD, M.A., F.R.S.

(American Editor, Professor J. McK. Cattell)

EARTHQUAKES

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EARTHQUAKES

IN THE LIGHT OF THE NEW SEISMOLOGY

BY

CLARENCE EDWARD DUTTON, MAJOR U.S.A.

AUTHOR OF "THE HIGH PLATEAUS OF UTAH," "HAWAIIAN VOLCANOES"
"THE CHARLESTON EARTHQUAKE," ETC.

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PREFACE

THE methods of studying earthquakes which have been developed during the last thirty years differ so much from those which preceded them, that they have justified the name of "The New Seismology." Prior to 1870, the studies published upon this subject were, with few exceptions, little more than narratives of disasters. Prof. Alexis Perrey had, it is true, compiled some extensive catalogues of earthquakes and published some comparisons which were interesting and which suggested inferences concerning the relations between earthquakes and the phases of the moon. Dr. Robert Mallet had investigated, with a minuteness never before attempted, the severe Neapolitan earthquake of 1857, and drew from his studies some remarkable conclusions. Neither Perrey's nor Mallet's work, however, added much of permanent value to the science except the truly scientific purpose and spirit which animated them. In general, the seismology of more than thirty years ago was descriptive only, and might be regarded as just within the outermost pale of scientific philosophy. The new seismology is eminently scientific, and in the strictest sense, for it investigates its phenomena by means of instruments which measure force and motions, speeds and acceleration. Its field is within the great department of physics, and is a

part of the branch which treats of elasticity and wave-motion.

The standpoint of the new seismology is very different from that of the old. The latter often seemed to regard earthquakes as being one of the formative geologic forces of nature, whose origin was mysterious, but which accomplished important structural results. The new view treats them as pure effects of geologic forces, just as thunder is an effect of the electric discharge, and not the cause of it. As sound is merely the elastic vibration of the air, so, in the view of the new seismology, the earthquake is merely the elastic vibration of the earth-mass. Hence the science becomes in a great measure the investigation of elastic wave-motion in a solid medium. This medium is the earth itself. And since the modes and configurations of this motion depend almost wholly upon the nature of the medium which transmits them, they become a means of investigating some important problems relating to the condition of the earth's interior.

The new seismology may be said to have begun its work with the invention of the seismograph. There are many forms of seismograph, and many investigators have contributed to their development, so many, indeed, that it might seem invidious to single out any one of them. Yet it is only just to mention Prof. James A. Ewing, formerly of the University of Tokio, and now of Cambridge, as the first designer of the most important features of the principal seismographs used for analysing earthquake motion and recording it in conjunction with a time scale. By means of these instruments it became possible to inaugurate an en-

tirely new series of investigations of the nature of earthquake motion, and ultimately to create a new branch of science.

In this line of research a leader was necessary; one who was well versed in physical science, having a high faculty of initiative, combined with a candid temperament, and one who was willing to make the study a life-work. For the science is a profound one, and could expect little progress from the fitful zeal and capricious attentions of those who might be momentarily attracted to it and quickly repelled by failure to achieve results. Such a leader was found in John Milne. In the earliest stages of the new science, when it was furnishing more questions than answers, he had the field mostly to himself, but when the results began to appear and to multiply, chiefly through his patience and industry, new workers entered the field, and many of them were of a high order of ability. At present they are quite numerous, and Japan, Germany, Austria, Italy, and England have many learned investigators. Their combined efforts have made the science very instructive and interesting.

The following chapters are intended to summarise their more important results: Chapter I sets forth the nature of an earthquake according to the modern concepts. It defines the technical terms used in discussion, and describes the action taking place on the surface of the ground during a quake of great energy. Chapter II is a general discussion of the causes of earthquakes. Two causes are recognised, apparently quite distinct, though possibly they may have interrelations not yet recognised. The first cause is

volcanic; the second is that force which is presumed to be always active in disturbing the rocks which form the outer shell of the earth, resulting in the building of mountains, the folding or shearing of the strata, and the elevation and depression of the earth's crust. Thus we have two groups of quakes: the volcanic and the tectonic. They have in many cases distinct characteristics, and these are described in Chapters III and IV.

Two chapters, V and VI, describe in detail the more important instruments used in seismometry: (1) the best forms of seismoscopes; (2) the higher and more elaborate forms of seismographs, which analyse seismic motion and resolve it into its three co-ordinates; (3) the large Italian vertical pendulums; (4) the most approved forms of the horizontal pendulum for recording the disturbances originating at far-distant parts of the world.

Chapter VII discusses the details of seismic vibratory motion and explains the four kinds of waves with which the inquiry deals. Two of these kinds of wave, the normal and transverse, pass through the earth-mass. The third kind appears to travel along great circles around the earth. The definite establishment of this latter form of wave-motion is a surprise to seismologists, and is not as yet wholly free of mystery, though its reality no longer seems to admit of any question. The fourth form is seen only in the epifocal regions of great earthquakes, and does not appear to consist of elastic waves at all, but of secondary effects of other vibrations. They, too, have long been difficult to explain, and, though attested by overwhelming evidence, are still far from being fully understood.

The kinetic elements of harmonic or vibratory motion are space and time,—*i. e.*, the distance traversed and the time of traversing it,—in other words, the amplitude and period of vibration. To this subject a chapter is given. The amplitudes and periods are measured by the seismograph, which thus gives the measured data upon which all kinetic estimates are based. The relations of horizontal and vertical vibration, as disclosed by the seismograph, are briefly set forth.

Passing then to the kinetic aspects of seismic vibration, the subject of intensity is treated in two chapters. For the moment we may regard intensity as the degree of vigour with which the earth, and objects upon its surface, are shaken. Since seismographs are few and far between, there is ordinarily no other way to reckon intensity than by its effects upon structures or loose objects, or upon human sensations. These effects are classified into a conventional scale of intensity ordinarily having ten degrees, ranging from the feeblest sensation of which men may be directly conscious to the energetic shaking which lays a city in ruins. The two most common scales in use, the Rossi-Forel and the Mercalli, are given. It might seem as if this way of reckoning intensity were exceedingly rough guesswork, of little value as a basis of kinetic estimates; yet nothing could illustrate better the highly scientific character of the new seismology than the attempts of Professor Holden and Professor Omori to correlate these purely conventional scales with an absolute scale of intensity, or accelerations.

The chapter on variations of intensity points out the method of computing the depth of origin of an earthquake

whenever observations sufficient in number and accuracy can be obtained. Other considerations relative to depth of origin are also discussed.

The speed of propagation of seismic vibrations is then treated. No specific problem in connection with earthquakes has been more diligently investigated, and few are so difficult as this. It is only very recently that definite results upon this question have been reached. The chief trouble has been the great complexity of the waves generated by an earthquake, their different rates of propagation, and the difficulty of separating one kind from another. Nor was it known until recently that some kinds of waves are propagated through the earth-mass, while others go around it.

Since the speed of propagation depends wholly upon the ratio of elasticity to density, it becomes an index of those properties in the materials which compose the earth's interior. Chapter XIII is given to the discussion of this aspect of the subject. It includes a table of elasticities and rigidities of rocks as measured by Professor Nagaoka, of the University of Tokio. It has a special value because all tables of solid elasticity determined in Europe have related only to metals, wood, minerals, and structural materials, while Professor Nagaoka's relate to rocks as they occur in nature.

The subject of earthquake distribution, or seismic geography, is treated in two chapters. In this field the science owes much to Major De Montessus de Ballore, who has gathered together a catalogue, in comparison with which those of Perrey, Mallet, and all predecessors look small.

His correlation of quakes with regions of strongly accentuated topography, or great topographic slopes, is a generalisation of capital importance. His tables of "seismicity," derived from the comparison of more than 130,000 quakes, are deemed of such importance and interest that they are given in an appendix.

The final chapter is devoted to the discussion of sea-quakes, a subject which has been investigated with great diligence by Dr. Emil Rudolph, of Strassburg. As the result of his researches, the discovery of several regions of great seismic activity in the ocean is of interest.

Although the greater part of the subjects herein summarised lies within those branches of physics which are usually treated mathematically, the effort has been to bring them within the range of popular science. Mathematical forms have been generally avoided. They have been employed only in a very few cases, when accurate expression was otherwise impossible.

C. E. D.

ENGLEWOOD, N. J., April 26, 1904.

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THE subject of earthquakes naturally divides itself into two lines of inquiry. The first relates to what occurs upon the surface of the ground, the other to what takes place below it. The surface actions, being seen and felt, are matters of direct observation, while the subterranean action is altogether a matter of inference. But while pursuing either line of inquiry we can never be unmindful of the other, or else we should cease to view the phenomenon in its entirety. We are compelled to regard an earthquake as having its origin or cause beneath the surface, while the movements that are seen and felt are regarded as effects transmitted to the surface from the origin.

The typical idea of an earthquake as a whole involves the concept of a central point below the surface, from which the vibratory impulses originate, or are propagated in all directions, like sound-waves, or waves of light. This point of origin is called the *centrum*. It is ideal to a large extent and never strictly real, though it seems essential to consistent reasoning. Yet the real facts are believed to constitute approximations to the idea sufficiently near to serve the general purpose of the truth without material error.

It is clear that the centrum cannot be a literal point; it must be a *locus* occupying a space with three finite dimensions. Its shape may be anything. In truth there are many earthquakes which we have reason to think have been caused by those shearing movements of the rocks which produce what geology terms faults. Others are known which appear to have been caused by the sudden dropping, or fall, of a considerable tract of the earth's surface with its underlying rocks, of great but unknown thickness. In such cases large areas of the surface, more than a hundred square miles in extent, suddenly sank, causing the earth to tremble many hundreds of miles away.

But though such movements seem at first to have little connection with the notion of a centrum, further reflection will indicate a certain consistency with it. For size is relative and the depth of the centrum is not fixed. A tract of three or four hundred square miles is but a small spot in one of tens of thousands of square miles, and, being always centrally situated with respect to the shaken area, may be regarded as the centrum. Nor is this consistency seriously affected by the fact that the sunken block has for its upper

surface the surface of the ground, for its vertical dimension is doubtless considerable and the portion whose impact generates the tremors is far below.

The idea of a centrum, therefore, may be taken with some qualifications, none of which materially affect its validity. It is a locus but not a point. It may be at any depth, of any size, and of any shape. A wave, or series of waves, is generated upon the still surface of a pond, whether we throw into it a crooked stick at any angle, or whether we drop into it vertically a perfectly spherical pebble. Though there will be differences in the minor details of the resulting waves, the effects are generically the same. So, too, with the generation of earthquake waves. Indefiniteness and irregularity of locus of origin is of consequence only in the details of the resulting waves, not in their general nature.

Wave-motion is such a complex subject and the waves which actually occur in nature are so varied in kind and so vast in number that it is a sufficient burden upon the mind and attention to follow out the action of a single wave of the simplest kind with as much elimination of complexities as possible. Limitations of speech and expression, therefore, compel us to treat of the centrum as if it were a literal point from which a single simple wave, or series of successive harmonic waves, emanated. Otherwise we should be lost in the complexities of the subject. The centrum, therefore, becomes an ideal point and not a real one.

A point on the surface of the ground, vertically over the centrum, is termed the *epicentrum.* This, too, is an ideal point and one that can never be determined with exactitude,

though it can be often designated with sufficient approximation. It is of very great utility in grasping the phenomena of an earthquake, since the surface effects are most pronounced at that point and fade out in all horizontal directions as we recede from it. The vertical line joining the centrum and epicentrum is called the *seismic vertical.*

The degree of vigour with which any point reached by the quake is shaken is termed the *intensity.* The term is rather a vague one when brought to the rigorous test of physical definitions, for it is the product of several factors, and its real meaning in use and as applied to facts on the ground is often limited to one of those factors rather than to their product. But in a general sense it is a good one, and is all the more convenient in many cases by reason of that very vagueness, or, rather, flexibility. It is generally employed to express the accelerating force acting upon a vibrating particle or element of the ground.

On the surface of the ground the intensity is greatest at the epicentrum and diminishes in all directions as we pass away from it. If the intensity faded out at the same rate in all directions, all points at equal distances from it would experience equal intensity. As a matter of fact, it diminishes unequally in different directions, so that points of equal intensity are farther away from the epicentrum in some directions than in others. If a line be drawn through points of equal intensity it will be found to be a closed curve around the epicentrum, more or less irregular in shape, often a deformed ellipse, sometimes, though seldom, roughly approaching a circle. Such lines are termed *isoseismals.*

The action being most violent in the near vicinity of the

epicentrum and fading out as the distance from it increases, it is sometimes convenient to have a term expressing the area which is most forcibly shaken, as distinct from the entire region throughout which it is felt. For this purpose Mallet employed the term *meizoseismic area*. It has no sharp definition, since the decline of energy with distance is gradual, according to a law which will be discussed hereafter. There is, however, a more rapid rate of decline along a certain part of the radius extending away from the epicentrum, and the location of this part depends upon the depth of the centrum. It may be taken somewhat arbitrarily as the limiting distance from the epicentrum, within which the shaking may be regarded as meizoseismic. Since it will be found in every direction from the centrum an isoseismal may be drawn through those points where the decline is most rapid, and to this ~~particular isoseismal~~ Seebach gave the name of *pleistoseist*. It has occasionally some convenience.

The rate at which the vibrations are propagated along the surface appears to vary in different earthquakes. Whatever may be the speed there will be some closed circuit around the epicentrum, making a line at which the shock of a single impulse may be expected to arrive at the same instant of time. Lines or curves drawn through such points are called *coseismals*. When the scientific study of earthquakes began, much was hoped from the use of them, but the hopes have not been realised. A great number of time observations is necessary in order to establish them, and these are seldom made. Coseismals, therefore, are seldom obtainable.

For the present it is assumed that the reader is acquainted with certain elementary features of that form of motion

called wave-motion, the fuller discussion of which is postponed to special chapters. Here it must suffice to say, that in solid bodies we distinguish two classes of waves, the normal and the transverse. In the normal wave the vibration of a particle affected by it is in the direction of a radius from the centre or origin of the wave; in the transverse wave the vibration is across or around that radius.

With these few definitions we may proceed to set forth in the simplest and most typical form the general idea of an earthquake as a whole. Of the subterranean causes which generate them we cannot here make any special postulates beyond the assumption that in the depths of the earth there occurs a sudden movement, or displacement, of matter in some tract or locus of considerable extent, as a result of which some strain is quickly generated or relaxed. This change of stress is the inception of a wave, and is propagated indefinitely through the earth in the manner of wave transmission. And not only one, but many waves of the same kind are generated. Each wave may (for simplicity's sake) be regarded as a spherical shell[1] surrounding the origin as a centre and expanding outward in all directions, like a sound-wave in the air. These waves at length reach the surface, where they impart motion to the soil and superficial rocks whose vibrations constitute the sensible earthquake. As the spherical wave expands, the intensity diminishes theoretically as the square of the distance from the centrum increases, really in all practical cases at a rate which is even

[1] The idea of a spherical shell containing between its inner and outer surfaces the locus and energy of the wave is not strictly correct, but no error sufficient to invalidate our present purpose will be incurred by regarding it, or a large part of it, in this way.

more rapid. If the earth-mass were perfectly homogeneous and elastic, the intensity would decline exactly in proportion to the square of the distance. But as it is far from perfect in either respect, a portion of the energy is gradually dissipated, and the decline is more rapid than the square of the distance. The action of a wave in an elastic solid as it reaches a boundary surface will receive attention in a future chapter. Here it can only be noted that the earthquake waves, after passing through hard rock-masses having elasticity, and reaching the looser and more heterogeneous materials at the surface, undergo transformations, and generate in those materials waves, vibrations, tremors, of a different order. The motions which are seen and felt at the surface are not the original vibratory waves rushing through the solid earth, which the type-theory contemplates, but are secondary movements resulting from them.

So considered the earthquake becomes a special case of radiant energy. This view was more or less dimly suspected or guessed before the nineteenth century, but appears to have received its first definite exposition from Dr. Thomas Young, to whom, in conjunction with Fresnel, the world owes so much for putting into workable form the undulatory theory of light. His views were afterwards amplified and developed by Gay Lussac. A great debt is also due to Robert Mallet, whose works on observational seismology and upon the Neapolitan earthquake of 1857 almost created a new department of physical science. He adopted the same concept of an earthquake and applied it with remarkable ingenuity to the actual phenomena. The advance of the science has in the meantime greatly modified Mallet's

methods and conclusions, but the powerful impetus he gave to the study marks his work as an epoch-making one in seismology.

The vibratory character of the phenomenon, however, had always been recognised. The old Greeks, though having no conception of elastic waves nor of the modern ideas of elasticity, fully realised the vibratory character of the disturbance. This is manifest in the brief treatment given the subject by Aristotle and Pliny. In the work *De Mundo*, Aristotle classifies the different movements, and distinguishes clearly enough between the duration of a single oscillation and the duration of the quake as a collective event comprising numerous oscillations.

Until within a comparatively recent period it has been customary to speak of an earthquake as causing the upheaval or downthrow of mountains and large tracts of territory. Thus Dr. Hooke, a contemporary of Newton, had classified earthquakes according to the effects they were supposed to have produced. One class caused elevations of the land; another caused depressions; a third caused horizontal displacements. The elastic-wave hypothesis reversed this relation and made the earthquake an incidental effect of the force which raises or lowers the ground, or of any force which may suddenly call into action the elasticity of the earth-mass. The old habit, however, of attributing great effects to seismic action still reveals itself occasionally even to-day, and we sometimes find men of deservedly high scientific repute inadvertently dropping into the practice of so regarding them. A similar misconception is still more common in popular writings.

The elastic-wave concept opens the way towards an explanation of the causes of earthquakes. Anything which calls into sudden action the elasticity of the earth-mass causes an earthquake, whether it be the fall of an acorn or the dropping down of a mountain range along a fault-line; whether it be the blast in a tunnel or quarry, or the outbreak of a volcano. The fundamental idea is of something which sets the elastic earth vibrating.

Importance is attached to the requirement that the elastic action be *suddenly* evoked. By the word "sudden" is not meant a matter of hours or even of minutes, but of a second or two at most, and much more frequently of a small fraction of a second, like the time involved in an impact, or a blow, or a quick collapse, or an explosion, or the sudden snapping under an elastic strain. This sudden action is essential to the production of a *sensible* vibration. Slowly acting forces can generate only correspondingly slow vibrations which become less sensible the longer their periods. We should be as unconscious of an earth oscillation occupying five or ten minutes as the passengers in midocean are of the rise and fall of the tides.

Theoretically there are no assignable limits to the greatness or smallness of earthquakes. In the direction of smallness, however, it is customary to draw the line at those quakes which are just forcible enough to be faintly felt and which are presumed to have the same origin and general nature as the larger or stronger ones. But the distinction is purely arbitrary and a mere matter of convenience. Modern research has assumed that below the scale of unaided sensibility the earth is pervaded with almost constant

minute tremors. What portion of them represents the unrest of the earth beneath and what proportion emanates from surface action—the fall of waves on the beach, the plunge of the cataract, the roll of the railway train—we do not know. But that many of them are generically earthquakes does not seem doubtful. They have in recent years been the subject of investigation by the Italian seismologists under the lead of Professors Bertelli and di Rossi. They are termed *microseismic tremors*.

We may now pass to a brief contemplation of those actions which take place on the surface at the time of a great earthquake and which are matters of experience and observation. Just here the description may be very brief, leaving a somewhat more detailed description to future chapters.

Great earthquakes come without any intelligible warning. It is true that such catastrophes have been preceded in a considerable number of instances by minor shocks and quivers and by ominous sounds. But these are far from implying, necessarily, a subsequent disaster, for they occur a hundred times without further consequence. It is only after the great shake that the mind recurs to them as its forerunners. Judging after the event they may indeed be regarded as its precursors. Judging before the event it is highly improbable that they will prove to be so, but not impossible. But the minds of men who have passed through such an ordeal are prone to seek premonitions in every circumstance which can be so interpreted, and the more imaginative ones think they have found them without sufficient reason for thinking so. Much has been written about

PLATE I.

Wreck in the Great Japanese Earthquake of October 31, 1891.

"earthquake weather," about a certain indescribable electric condition of the atmosphere, about the sensitive and alarmed conditions of animals, about the erratic flights and actions of birds, just before an earthquake. If any such phenomena really manifest themselves as preliminaries to an earthquake all that can be said about them is that they are as mysterious to the seismologist as to everybody else. But the testimony

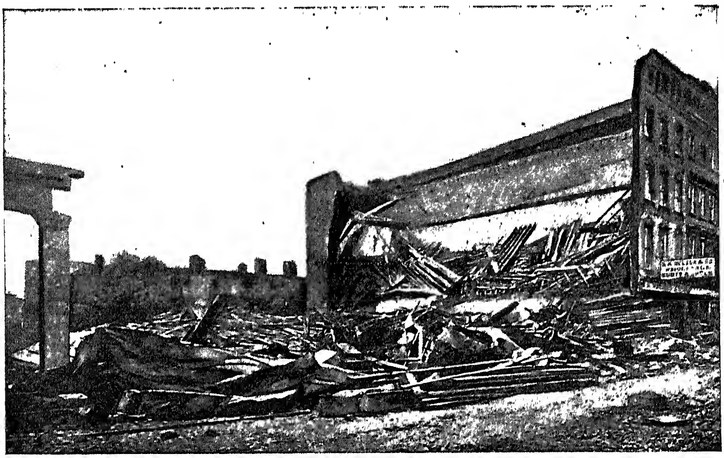


FIG. 1. Destruction in Hayne Street, Charleston, S. C., by the Earthquake of Aug. 31, 1886.

in support of them, though considerable in amount, is vague in character. When carefully scrutinised it leaves the impression that it is the outcome of the imagination and not of real observation, or that very commonplace facts have been given characters and relations which are not warranted, and that to whatever extent the facts may have been real they had no more to do with earthquakes than the changes of the moon have to do with the weather.

When the great earthquake comes, it comes quickly and is quickly gone. Its duration is generally a matter of seconds rather than of minutes, though instances have been in which it lasted from three to four minutes. Perhaps forty-five seconds would be a fair average. The first sensation is a confused murmuring sound of a strange and even weird character. Almost simultaneously loose objects begin

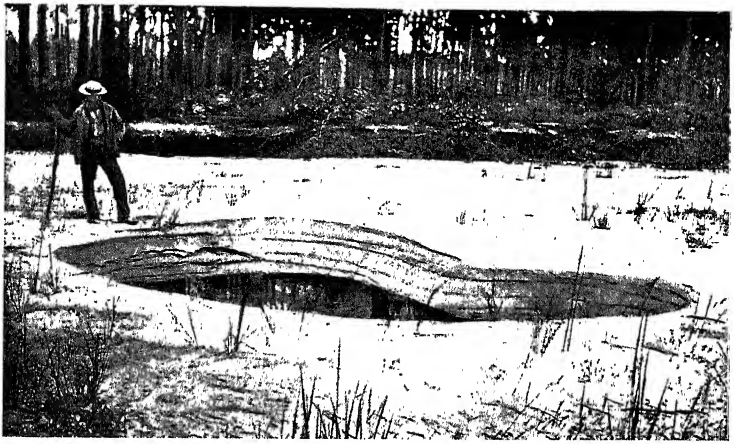


FIG. 2. Craterlet Formed in Charleston Quake from which Sand and Water Were Discharged. There Were Many Hundreds of Such Craterlets near Summerville.

to tremble and chatter. Sometimes, almost in an instant, sometimes more gradually, but always quickly, the sound becomes a roar, the chattering becomes a crashing. The rapid quiver grows into a rude, violent shaking of increasing amplitude. Everything beneath seems beaten with rapid blows of measureless power. Loose objects begin to fly about; those that are lightly hung break from their fastenings. The shaking increases in violence. The floor begins

to heave and rock like a boat on the waves. The plastering falls, the walls crack, the chimneys go crashing down, everything moves, heaves, tosses. Huge waves seem to rush under the foundations with the swiftness of a gale. The swing now becomes longer and still more powerful. The walls crack open. A sudden lurch throws out the front wall into the street, or tears off or shakes down in rubble the whole corner of the building. Then comes a longer swaying motion, not like a ship at sea, but more rapid; not alone from side to side, but forward and backward as well, and both motions combined into a wriggle which it seems impossible for anything to withstand. It is this compound, figure-8 motion which is so destructive, rending asunder the strongest structures as if they were adobe. It is the culmination of the quake. It settles into a more regular swing of decreasing amplitude, then suddenly abates and the motions cease.

Or suppose we are out in the country and the earthquake comes suddenly upon us. The first sensation is the sound. It is wholly unlike anything we have ever heard before unless we have already had a similar experience. It is a strange murmur. Some liken it to the sighing of pine trees in the wind, or to falling rain; others to the distant roar of the surf; others to the far-off rumble of the railway train; others to distant thunder. It grows louder. The earth begins to quiver, then to shake rudely. Soon the ground begins to heave. Then it is actually seen to be traversed by visible waves somewhat like waves at sea, but of less height and moving much more swiftly. The sound becomes a roar. It is difficult to stand, and at length it becomes impossible

to do so. The victim flings himself to the ground to avoid being dashed to it, or he clings to a convenient sapling, or fence-post, to avoid being overthrown. The trees are seen to sway sometimes through large arcs, and are said, doubtless with exaggeration, to touch the ground with their branches, first on one side, then on the other. As the

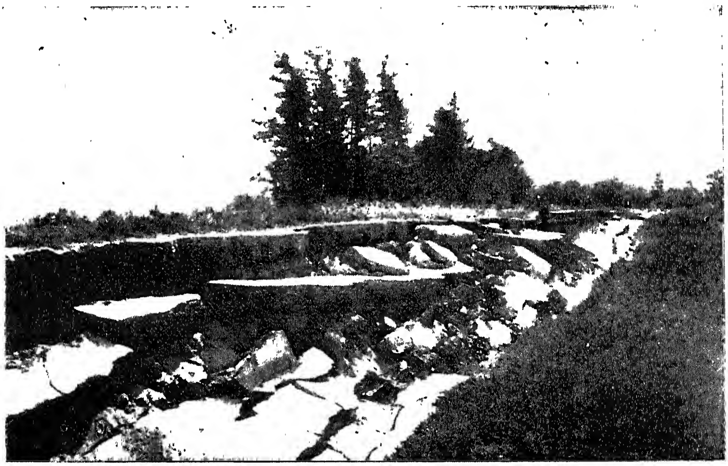


FIG. 3. Landslip, or Sliding of a River Bank towards the Stream, a Common Occurrence in a Great Quake. Mino-Owari Earthquake.

waves rush past, the ground on the crests opens in cracks which close again in the troughs. As they close, the squeezed-out air blows out sand and gravel, and sometimes sand and water are spurted high in air. The roar becomes appalling. Through its din are heard loud, deep, solemn booms that seem like the voice of the Eternal One, speaking out of the depths of the universe. Suddenly this storm subsides, the earth comes speedily to rest and all is over.

And yet this feeble description suggests but a single instance, or a few instances having a general similarity. There are many variations of detail in the incidents of great earthquakes. In some the full vigour of the shock comes without any *crescendo*, but as a *sforzando*, with an almost

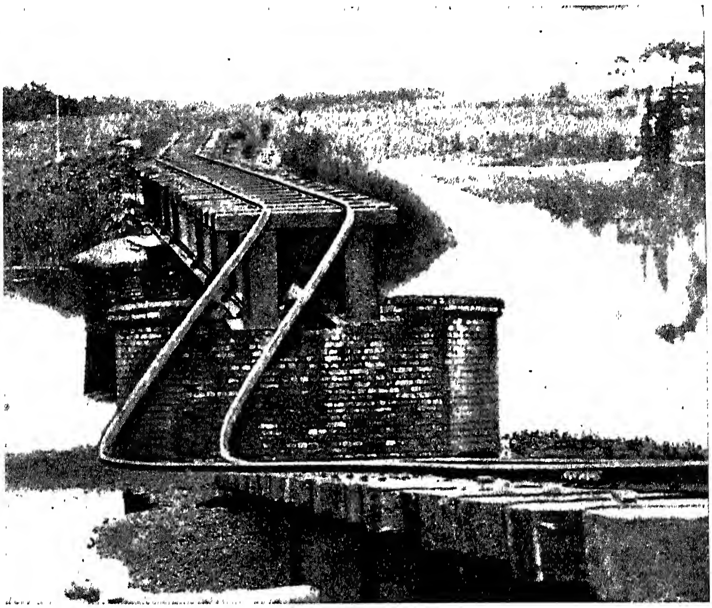


FIG. 4. Fall and Lateral Displacement of Railway Bridge and Pier in the Great Bengal-Assam Quake of July 12, 1897.

explosive suddenness. People find themselves suddenly thrown to the earth, the ground literally knocked from under their feet. Sometimes the rolling waves of soil are absent and the motion is a rude quiver, rapidly vibrating in every direction, twisting, contorting, wrenching the ground, as if in a determined effort to shake it into dust

Sometimes the most pronounced motion is vertical, as if the earth beneath were being hammered upward by a quick repetition of strokes. Sometimes the crescendo, climax, diminuendo are immediately repeated before the first cycle has come to complete rest, or it may be twice repeated. Or an interval of some minutes may elapse before the repetition, or even several hours. But a long-deferred repetition is very uncommon, though nearly all great earthquakes are followed by minor shocks for days, weeks, months, or even years afterward. Some of these are of considerable force though not of the devastating power of the principal shake. But they are alarming enough to keep people in a state of apprehension until they become inured to them. As time passes they diminish both in force and frequency and at length cease.

CHAPTER II

CAUSES OF EARTHQUAKES

The Three Views—The Dislocation Theory—The Volcanic Theory—Views of Humboldt and Boussingault—The Tidal Theory—Perrey's Inquiries and Laws—The Statistical Method of Investigation—Milne's Catalogue in Japan—De Montessus de Ballore—Sources of Earth Stresses—The Contractional Hypothesis—The Ideas of Babbage and Herschel—Isostatic Equilibrium—Prof. George Darwin's Discussion of Stress-differences and their Distribution—The Relative Magnitudes

THE originating or ultimate causes of earthquakes have been the subjects of controversy for more than a century. It would be curious and amusing, rather than useful, to go back into the eighteenth century to exhume buried and forgotten hypotheses put forth to explain these phenomena. They seem to indicate the fanciful childhood of the science which endeavours to deal with problems of earth physics. It will be more profitable to limit ourselves to those explanations which, though they may be more or less erroneous, are neither childish nor absurd.

During the first half of the nineteenth century these might be divided into three groups: 1st, those which attributed earthquakes to sudden downthrows or collapses of the ground; 2nd, those which attributed them to volcanic action; 3rd, those which attributed them to the

action of a liquid interior of the earth upon an external rocky crust under the disturbing influence of tidal forces. As none of these theories have been fully disposed of, and as all of them still have more or less vitality at the present time, we may in a preliminary way briefly advert to them.

The downthrow, or downfall, or *Einsturztheorie* has had for its advocates J. J. Scheuchzer, in his day an eminent Swiss geologist[1]; Boussingault,[2] a far more illustrious naturalist; Albert Necker,[3] and G. H. Otto Volger.[4] These authors were disposed to give the largest possible extension to the downfall theory. Such downthrows, instantly followed by great and far-reaching earthquakes, had been witnessed and verified in the calamities of Port Royal in Jamaica, 1692, in the terrible Calabrian quake of 1783, in the New Madrid disaster, 1811–12, in the Rann of Cutch near the mouth of the Indus, 1819, and in Murcia in 1829. The studies of Boussingault, Humboldt, and Darwin in the Chilian, Peruvian, and still more northerly Andes had established the occurrence of many others under similar conditions.

That a considerable number of the greatest and most

[1] Scheuchzer unfortunately is remembered at the present time chiefly through a blunder he made in describing the fossil bones of a large lizard as human remains, and as *homo diluvii testis*. But this does him a great injustice, for he was an acute observer.

[2] "Sur les tremblements de terre des Andes." *Annales de Chimie et Physique*, vol. lviii., 1835.

[3] "On a Probable Cause of Certain Earthquakes." *Phil. Magazine*, vol. iv., 1839.

[4] "Untersuchungen über das Phänomen der Erdbeben in der Schweitz." Gotha, 1857–58, drei Bande.

The above references are taken from Dr. Rudolf Hoernes's *Erdbebenkunde*.

forcible of modern earthquakes were attributable to sudden displacements in the rocks seems to be well established. The chief error of the *Einsturztheorie* seems to have been the extreme lengths to which its advocates carried it. Not content with claiming its validity in those cases where the sudden downthrow was well attested and still open to the inquest of the observer, with the instantaneous sequence of the earthquake proven by unquestionable evidence, it was sought to apply the same action to quakes in which no evidences of downthrow could be pointed to. It was by many thought necessary to postulate large subterranean cavities into which great masses of overlying rock might drop without the least sign upon the surface that any such cavity had existed, or that any such drop had occurred.

The more recent views of downthrows in the outer shell of the earth do not require the pre-existence of large subterranean cavities into which great blocks or masses of the earth may suddenly fall. They are regarded rather as the results of unequal stress in the outer shell which have become so great that the rocks must yield to them along lines of least resistance as if they were plastic instead of rigid. The masses involved seek a readjustment or equilibrium of stress, and the motion may be upward in one locality and downward in an adjacent one, and probably also compounded with horizontal shifting in either case. At no time and at no place is any cavity supposed to exist any more than a cavity under the water surface is required to explain the sinking of the wave-crest into the wave-trough.

In those instances where the downthrow had surely occurred the inquiring mind was left in the utmost perplexity

to imagine how a vacuity large enough to receive the fallen mass could have existed. Caves near the surface, usually subterranean watercourses dissolved out of limestones, are about the only vacuities in the earth of which we have positive knowledge. But they are insignificant in size compared with vacuities which would be necessary to accommodate the masses involved in such downthrows as those of New Madrid, Murcia, or the Rann of Cutch. The deeper the imagination penetrates into the earth the greater becomes the pressure of the overlying rocks, and the greater the improbability of any cavity. The disputant who postulates them must be required to bear a heavy burden of proof. The downthrow theory cannot claim full acceptance beyond those instances where the evidence of the downthrow is patent to every eye upon the surface of the ground and when the instantaneous sequence of the earthquake is attested by satisfactory evidence.

One other error was made by the advocates of this theory, and that was the endeavour to discredit all other theories unduly, especially the volcanic theory. At that time (the first half of the nineteenth century) controversies among geologists and zoölogists were very bitter. It seemed as though the *odium theologicum* had taken possession of them and generated an asperity surpassing even that of clashing creeds and political prejudice. The volcanic theory was a rival to their own, and with a bias that at times was almost passionate they were apparently disposed to deny it any but the most insignificant part in the category of causation.

The belief that earthquakes were associated with volcanic action, that they were caused by it, or were in some way dependent upon it, is as old as Aristotle. It appears in the works of Pliny, Strabo, and Pausanias, and was universal throughout the Middle Ages. These opinions were, however, tinctured with more or less superstition and not a little absurdity until near the beginning of the nineteenth century. The volcanic theory first took on its scientific form under the treatment of two of the most illustrious naturalists of that period, Leopold von Buch, and Alexander von Humboldt. Curiously enough, both of them were pupils of the Neptunist Werner, and if early education could have biased minds of such calibre it must have tended to array them against a Plutonic source of causation. Both of them, however, held to the view that the earth's interior was composed of molten matter, and that volcanic action was the chief geotectonic agency. Humboldt, however, regarded volcanism as being the reaction between the hot interior and the cold exterior, and regarded volcanos as safety-valves. So long as they remained open the interior forces could not, he thought, accumulate to any dangerous extent in their immediate neighbourhood. But at a distance from the vent they might become more menacing. He seemed to be much impressed with an opinion which he found almost universal in the Andes among dwellers in the vicinity of volcanos, that so long as the crater continued to steam freely and to discharge regularly puffs of dust and lapilli there was no danger of serious developments. But if activity ceased at the upper cone there was reason for grave forebodings.

Humboldt was also impressed with a fact which Boussingault had already observed in the Andes and had made much use of in support of the downthrow theory. He found that such quakes as occurred near an open, active volcanic vent were almost always light and inconsiderable, while the heavy and destructive ones were almost always far away from them. This statement has proved to be even more general than Humboldt supposed. It is of world-wide application. The instances of heavy destroying earthquakes in close proximity to active volcanos are extremely few, though a very small number are recorded.

Opposite interpretations were given to this fact by the advocates of the *Einsturztheorie* on the one hand and the advocates of the volcanic theory on the other. The former regarded it as strong evidence that volcanos had little to do with earthquakes, and were an effective cause only of some minor ones. The latter regarded it as proving that earthquakes were generally caused by Plutonic forces whose energy and destructiveness were much tempered and reduced by the close proximity of an open volcanic vent which acted as a safety-valve.

Ever since the appearance of Humboldt's accounts of his journeys in South America and Mexico the public at large—not merely intelligent readers in general, but most scientific men who concern themselves with problems of terrestrial physics—have taken for granted the interrelation of volcanism and seismism. It is the almost universal belief that volcanos and earthquakes occur together, that the chief earthquake regions are volcanic, and *vice versa*. From this

belief the passage is easy and natural to the inference that the volcanic forces are the causes of earthquakes; if not of all of them, then of nearly all. This inference is strengthened by another consideration. Very few men of science, fewer still of other callings, ever trouble themselves to acquire a knowledge of any except the most memorable and destructive earthquakes. Of those which they have noted, the greater part have occurred in countries which abound in volcanos, active or extinct, and they pass easily to the idea that the two are in close proximity and, therefore, associated. A detailed examination in such cases shows that the earthquake centre is usually so far away from any volcano as to render any association between them highly improbable.

On the other hand, there have been some quakes originating so near to a volcano that a connection is suggested at once, and on being examined in detail the presumption becomes a practical certainty. Thus the repeated destruction of cities at the base of Mt. Ætna, the terrible quakes in Hawaii which immediately preceded the eruption of 1868, the calamity of Casamicciola on the island of Ischia in 1883, were all associated with volcanos in such a way as to leave no doubt that the kinetic cause of them was volcanic in its nature.

This suggests that a study of the distribution of volcanic vents and earthquakes in some locality where both occur would be instructive. This has been done for the Japanese empire by Prof. John Milne, and the result of his study will be given in the next chapter. Here it must suffice to say that the many quakes of that archipelago seemed to

show an avoidance of the volcanic centres which are numerous in the interior and to indicate that volcanic energy was seldom concerned in generating them.

Similar studies, though less systematic and exhaustive, have been made of the relations of earthquakes to volcanos in Italy. The result is that while the interrelations suggesting cause and effect are somewhat more frequent in Italy than in Japan, and doubtful associations considerably more frequent, yet it is surprising how much greater has been the number of quakes which suggest no such dependence. Finally, a widely extended study of the same question for all those regions of the world which have been the fields for seismic investigation has been undertaken by Capt. F. De Montessus de Ballore, and the general result is that not only is the proportion of quakes for which a volcanic origin is indicated a small one, but that it is much smaller in the great and powerful class than in the moderate and feeble ones.

It remains to glance briefly at the tidal theory. That the forces exerted upon the earth by the moon, and their variations through the different parts of her orbit, might, among other results, be effective in promoting earthquakes is an old idea.[1] It was first propounded in scientific, definite form, and supported by an array of tangible evidence worthy of serious consideration, by Alexis Perrey of Dijon. The foundation of Perrey's view was the prevalent belief that the earth's interior is in a state of fusion by reason of

[1] In *Comptes Rendus*, xxxix., 1850, p. 375, M. Zantedeschi refers to similar views of G. Baglioli in *Historia Romana terræ motus anni*, 1703. Opera omnia, pub. 1737; also to G. Toaldo, della vera onfluenza degli astri, Padua, 1770, in the discussion of Peruvian earthquakes.

aboriginal heat. Upon this liquid mass, enclosed by a thin, rocky crust, or skin, the tidal action of the moon continually exercised a disturbance which reacted upon the crust so as to produce cracks, fissures, and displacements, with earthquakes as accompaniments. Perrey had spent many years in compiling a great catalogue of earthquakes extending as far back as the fourth century A.D., and ultimately made a list of nearly twenty thousand of them. Unfortunately they have never been published in a collected form, but are scattered through the proceedings of several scientific societies. A list of these disjointed publications will be found in the Royal Society Catalogues. To Perrey is due the high credit of having originated the method of collating earthquake statistics and of studying them *en masse* in various relations—a method now known as the *statistical method*, which has recently been expanded by Milne in Japan, de Montessus De Ballore in France, C. W. C. Fuchs in Germany, and Charles Davison in England with highly valuable results, some of which we shall presently consider.

After amassing about fifteen thousand records of earthquakes, Perrey proceeded to study his catalogues to ascertain what light they could throw upon the action of the moon as a promoting cause. As the result of these studies he announced three laws which are still known as Perrey's laws. They are as follows:

I. Earthquakes are more frequent at the syzygies and less frequent at the quadratures.

II. They become more frequent as the moon approaches perigee and less frequent as it approaches apogee.

III. They are more frequent when the moon is near the meridian than when it is ninety degrees from it.

Perrey grouped his data in many ways, so as to cover a considerable number of periods and localities as well as to take larger groups of wider range. His process consisted in counting the number of lunar days by dividing the lunation (29.53 days) into eight equal parts and by combining the first and eighth, the fourth and fifth, for the syzygies, and the third and fourth and sixth and seventh for the quadratures. Taking, for example, the period 1801 to 1850 inclusive, he found 5388 lunar days on which earthquakes occurred. But it happened that on some days more than one quake occurred in widely separated localities. It was quite proper, therefore, to estimate in such cases each quake as a day, or to count such days more than once. This correction gave him 6596 lunar days. The 5388 days without duplication gave 2761.48 at syzygies, 2626.52 at quadratures, an excess of 134.96 or 2½ per cent. at the syzygies. The 6596 lunar days gave 3434.64 and 3161.36 respectively, or an excess of 273.28 or four per cent. at syzygies.

Again: Taking the earthquakes at Reggio in Calabria as given for the period 1836 to 1853 (eighteen years) in a journal kept by M. S. Areovito, he finds 437 at the syzygies and 349 at the quadratures, an excess of eighty-eight, or over eleven per cent. at the syzygies.

By similar methods he compared the number of quakes occurring at apogee and perigee and found a difference of six per cent. in favour of perigee.

By comparing the records kept at five Italian observa-

tories, he found the following relation between shocks and the position of the moon with respect to the meridian:

Observatory	Number of Shocks		Excess at Meridian	Percentage
	Less than 45° from Meridian	More than 45° from Meridian		
Monteleone, Jan., 1783, to Oct. 1, 1786	475	453	22	51.19
Messina, Feb. 5, 1783, to Jan. 2, 1784	84	60	24	58.33
Calanzuro, Feb. 5, 1783, to July 5, 1783	102	81	21	55.74
Scilla, Oct. 1, 1783, to Nov. 25, 1785	140	120	20	53.85
Reggio, Mar. 10, 1836, to April 7, 1853	413	347	66	54.35

Perrey's announcements awakened considerable interest and criticism.[1] These excesses at syzygies, at perigee, and at culmination, though small, were still large enough to merit consideration. But there was much reluctance to admit that so small a potential could produce such large results. M. Delaunay, the astronomer, though himself a believer in a liquid interior of the earth, regarded Perrey's views as resting on much too small a basis of observed facts. For though the total number of observations at his command was large it was necessary to consider many localities

[1] The *American Journal of Science*, vol. xxxvii., May, 1864, contains an article by Perrey setting forth his theory and the results of his statistical comparisons. *Vide* also *Comptes Rendus*, 1853, 1854, 1861.

in detail, and when the data were divided up among a considerable number of localities the quota of each would be small. And M. Perrey himself had shown that in some seismic districts the results would have been reversed.

Since Perrey's announcements the subject has been followed up by several investigators with a rapidly increasing amount of data. The number of recorded earthquakes which have been catalogued is now in the neighbourhood of 140,000, and the records of the newly acquired ones have a much greater scientific value than the old ones of Perrey's time. The result has been that as the number and scientific accuracy of the data have increased, the preponderances which Perrey found at syzygies, at perigee, and at lunar culmination have dwindled away to almost nothing, and in many localities have been quite reversed. Moreover, the theory of a molten interior of the earth has in England, America, and Italy been definitively abandoned in favour of an earth possessing a high degree of effective rigidity, and the same change of opinion is rapidly gaining ground in Germany. In France the cultivators of earth physics seem to be shaken in their belief in a liquid interior, though they are still very conservative in their views. With its fundamental postulate fast vanishing, with support derived from statistical comparisons dissolved away, little remains of the tidal theory of earthquake causation.

Thus far, then, we have two causes of earthquakes which are apparently well sustained: (1) the downthrows, which have often been observed to be accompanied by earthquakes, and (2) volcanic action. But neither of them have been shown to be connected with more than a compara-

tively small number. Much the greater part of the earthquakes still require explanation, and the indications are manifold that some of them are produced by some cause yet to be stated.

Dr. C. G. Knott very well remarks,[1] that "an earthquake always means a yielding to stress, whatever may have been the source of this stress." This highly general statement will probably receive the assent of nearly, if not quite, all seismologists. But it needs specification and analysis. In earthquakes which are clearly the accompaniments and presumable effects of volcanic action the same notion is implied. It is taken for granted that the subterranean movements of lavas forcing their way through or into the rocks, encountering resistance, generate stresses. It is equally taken for granted that downfalls of large masses of the strata generate sudden stresses of vast magnitude. Both of these modes of causation imply stress. But, inasmuch as the majority of earthquakes furnish no evidence either of volcanism or of downthrow, can we not find some other mode of stress which may account for them? Hitherto we have found only two modes which we can, with practical certainty, declare to be effective.

Two other modes of earth stress have been discussed which are of interest in this connection. The first is attributed to the secular cooling of the earth's interior and the slow, continuous readjustment of the cold, outer crust to the shrinking nucleus. This is an old view and it received a remarkable development at the hands of Dr. Robert Mallet in 1871.[2] It was wrought out with such ability and

[1] *Seism. Soc. Japan*, vol. ix., Pt. I., p. 2.

[2] *Phil. Trans.*

attractiveness that for some years it seemed to have gained the adhesion of most geologists and not a few physicists.

Subsequent criticism, however, showed[1]: (1) that if the earth ever was a uniformly heated globe, left in space to cool, the amount of cooling up to the present time has been inconsiderable, and below a depth of about one-fiftieth of the terrestrial radius is thus far negligible; in other words, the earth, excepting a thin outer shell, is as hot as it ever was, and that the interior, properly speaking, has not yet contracted at all; (2) that the corrugations, foldings, or crumplings of the strata are not such as would be produced by an outer crust continuously adjusting itself to a shrinking nucleus.

The contractional hypothesis is here regarded as being virtually destroyed by effective criticism, and in the form given it by Mallet as being simply an error.

There remains an older hypothesis originating from Babbage and Herschel, which takes account of the ultimate effects of that process which has been going on throughout the whole range of geological time, in which the materials derived from the disintegration of the rocks on the land are carried down by rivers to the sea or into the valleys and deposited there. This involves a shifting of loads from one part of the earth's surface to others, and as it is cumulative through the geological ages it must generate cumulative strains.

This fact must be considered in connection with the question whether the deformation of the terrestrial spheroid by the elevations which form the lands and the depressions

[1] *Physics of the Earth's Crust*, Rev. Osmond Fisher.

which form the sea basins are real deformations of the figure of equilibrium due to gravitation, or whether they are regions of less and greater density respectively. If they are in truth deformations of isostasy, or equilibrium of figure, they are already a source of great stress, and have been still more so in the past. For the process of denudation on land and sedimentation at sea is a continuous process of relief of that strain. Observations with the pendulum and with the plumb-line, however, furnish strong reasons for believing that they are not such deformations, but that the continents and plateaus with their greater mountain platforms are composed of, and deeply underlaid by, lighter matter than the sea bottoms, and stand higher because, so to speak, they float higher.

Geology furnishes also another fact of importance in this relation. It has been frequently observed that regions of great denudation, which are invariably lofty and mountainous, maintain their altitudes by positive uplifting about as fast as they are torn down and removed by secular denudation. We have also the complementary fact that regions of great sedimentation have, throughout the duration of that process, been shallow-water areas and have sunk as fast as the sediments were piled upon them. Our great Paleozoic system of the Appalachian region, 30,000 to 35,000 feet thick, was all deposited in shallow waters; for every stratum carries shallow-water fossils. Our Mesozoic system of Colorado and the Plateaus, 9000 to 11,000 feet thick, was also deposited in shallow water, for it contains at many horizons fossil forests alternating with beds carrying deinosaurs, lamellibranchs, and brachiopods. Our coast range

in California and Oregon, consisting of vast but unmeasured masses of Cenozoic beds, was also a shallow-water accumulation. On the other hand, the mountain ranges which supplied these western sediments are still standing, and it is not probable that they were ever any higher than they are now. Yet their present mass is but a small fraction of the mass of their own ruins which surrounds them.

With the geological aspects of this proposition we must concern ourselves as little as possible. What is of most importance to us is to point out that these transfers of material must set up stresses which in the lapse of indefinite time may become, unless relieved, almost indefinitely great, and that the geological facts indicate that the earth-mass has obeyed these stresses which in the last analysis are simply the consequences of terrestrial gravitation conserving the altitudes of the denuded lands and of the overloaded littorals or sea basins.

This isostatic conservation is thus conceived of as a flow of the earth-mass away from the areas which have been receiving great loads of sediments, and towards the areas which have been denuded, to supply them. This flow is resisted by the rigidity of the earth. But the masses involved are so great that no practical rigidity can withstand it. It is as resistless as the onward movement of a great glacier, or the Greenland ice-cap. It has two aspects: 1st, the outward pressure of the regions of sedimentation, and 2d the inward pressure towards the regions denuded.

These two phases of the action of terrestrial gravitation may sometimes conspire. A region of great denudation may send its detritus to the sea, where it is deposited near

the coast and not far distant from the highlands or mountains and plateaus which furnished it. The littoral then becomes a region or zone of loading, while the adjacent highlands are a region or zone of unloading; and the strain set up is the sum of the strains which each phase would generate. More frequently the detritus brought down to the sea is in greatest part distributed far and wide over the ocean, and the earth strain generated is mainly that which arises from the unloading alone.

In this brief presentation of the isostatic conception care should be taken to avoid confounding this equilibrating force with the force which uplifts or depresses the land in such a way as permanently to raise or lower the mean profiles of the land or sea bottom. Isostasy is the tendency to maintain profiles in equilibrium, not to raise or lower them. What are commonly spoken of as upheavals and depressions are due to some cause or causes not yet understood. Isostasy is merely one of the conditions under which such forces act. It neither helps nor hinders their action, and its own action is independent of them. Its whole scope is the conservation of the profiles after they are made, and not to make or change them. The forces of elevation and depression are of a totally different category and are yet to be discovered and explained.

Prof. George Darwin has discussed the subject of earth strains arising from unequal distribution of loads upon the surface in a profound and remarkable analysis published in the *Proceedings of the Royal Society* (June, 1881). In this discussion Professor Darwin assumes for the sake of argument that the land elevations and marine depressions are

real departures from a figure of equilibrium (isostasy), and that at equal distances from the centre the density of the earth is everywhere the same. Hence the interior of the earth must be in a state of stress, and as the land does not sink in, nor the sea-bed rise up, the materials of which the earth is made must be strong enough to bear this stress. He then proceeds to inquire how the stresses are distributed in the earth's mass, and what are the magnitudes of the stresses.

He estimates the strength of an elastic solid by the difference between the greatest and least principal stresses when it is on the point of breaking, or, according to the phraseology accepted, by the breaking "stress-difference" which is expressed in tons per square inch.

Professor Darwin solves the problem for a class of surface inequalities called zonal harmonics, which consist of elevations and valleys running zonally around the earth, like parallels of latitude, forming what might be conceived of as zonal corrugations. The number of corrugations or waves is determined by the order of the harmonics, and each order becomes the subject of special treatment.[1] In the application to a spheroid the equator may be any great circle, and, in the case of the earth, not necessarily the terrestrial equator.

The second harmonic has only a single wave and consists of an elevation at the equator and a depression at the poles, constituting an oblate ellipsoid. An harmonic of a higher

[1] The necessity for this will be apparent to mathematicians who are familiar with the subject of spherical harmonics. The different orders are even-numbered for equally obvious reasons.

order may be conceived of as a series of mountain chains, with intervening valleys running around the globe parallel to the equator. The case of the second harmonic is considered in detail, and it is shown that in this case the stress-difference has its maximum at the centre of the globe and is constant all over the surface, and that it is eight times greater at the centre than at the surface. If, then, we compute the stress-difference due to an excess or defect of ellipticity the one-thousandth part above or below the equilibrium value in a spheroid of the size and density of the earth, it would be about eight tons to the square inch at the centre. Or if a homogeneous earth with an ellipticity of $\frac{1}{232}$ were to stop rotating, the central stress-difference would be thirty-three tons per square inch, and it would rupture if composed of a material weaker than hard steel.

The stresses produced by harmonic irregularities of higher orders are then considered. It is found that the stress-difference depends only upon the depth below the mean surface, and is independent of the position of the point relative to the crests and troughs of the corrugations. Numerical solutions show that with a series of mountains whose crests are four thousand metres above the adjoining valleys, with rocks of 2.8 specific gravity, the maximum stress-difference is 2.6 tons per square inch, and if the mountain chains are 314 miles apart the maximum stress-difference is reached fifty miles below the mean surface.

The cases of harmonics of the fourth, sixth, eighth, tenth, and twelfth orders are then considered. It is shown that, if we suppose them to exist on a sphere of the size and mean density of the earth, and that the height of the elevation at

the equator in each case is 1500 metres above the mean level of the sphere, then in each case the maximum stress-difference would be about four tons per square inch. This maximum is reached in the case of the fourth harmonic at a depth of 1150 miles; in the case of the twelfth harmonic at a depth of 350 miles below the surface.

Finally, this shows that the greater inequalities of the earth, Africa, the Atlantic Ocean, the Americas, may correspond to, or be represented by, an harmonic of the fourth order in which the maximum stress-difference is about four tons per square inch occurring about 1150 miles below the surface.

This thesis of Professor Darwin enables us to form some conception of the stresses which would be generated by surface inequalities or transfers of load under conditions which specify masses, densities, and their distribution, together with the distribution of the rigidity by which such stresses are sustained.

With respect to rigidity and its distribution he makes no particular assumption, since the final object of his analysis is to find what indications the general analysis of earth-stresses can furnish as to the general theory of a highly rigid earth-mass propounded by Hopkins and Lord Kelvin. In other words, rigidity is the end of his inquiry, and not its beginning.

If, however, any considerable portion of the earth's interior be liquid or markedly viscous (*i. e.*, having a low degree of rigidity, say as low a that of lead) it profoundly modifies the amount and distribution of the resulting stresses. If, as some able geologists suppose, the earth

has a rigid and comparatively thin outer shell or crust and a highly rigid nucleus with an intervening layer of less rigid or slightly viscous matter, the stress-differences would be much increased near the surface and much diminished in the depths.

In the transfer of materials from one portion of the earth's surface to another a cause of stresses and of stress-differences appears, which is cumulative through long periods of geological time, and which may become great enough to cause sudden yielding of the rocky strata. That there may be other causes of stress is not denied. We have indications of more or less unrest of the earth's interior which we find it hard to explain. Volcanic action itself is not yet understood and we have no satisfactory theory of it. But the transfer of sediments is the only obvious and plainly visible cause which has thus far been suggested as the source of those cumulative stresses which ultimately become resistless and lead to the collapse which generates the earthquake.

CHAPTER III

QUAKES OF VOLCANIC ORIGIN

The Fire Girdle of the Pacific—Its Discontinuous Character—Exaggerated Estimates of the Interdependence of Volcanic and Seismic Activity—Milne's Japanese Catalogue—It Shows Little of Such Association—Milne's and De Montessus's Conclusions on the Association of Seismicity with Topographic Relief—Many Quakes are Certainly of Volcanic Origin—Distinctive Characteristics of Volcanic Quakes—Their Shallow Centra—The Casamicciola Quake of July 28, 1883—The Mauna Loa Quake of March 27, 1868—The Quake on Mt. Ararat, June 20, 1840—Recent Eruptions in Martinique and St. Vincent Unaccompanied by Marked Seismic Action—Volcanic Tremors—The Krakatoa Eruption—Small Relative Energy of Volcanic Quakes—Quakes from Mt. Ætna—Possibility of Quakes from Plutonic Subterranean Action Not Manifested on the Surface

MUCH has been said in elementary books of Geology and Physical Geography of the great volcanic circle which, on the east of the Pacific, extends from Alaska to Cape Horn, and on the west of that ocean extends along the border of Asia and its archipelagos to the equator, and thence westward to the Bay of Bengal and southward to New Zealand. On a globe a foot or two in diameter this girdle appears as a nearly continuous belt, for the most part narrow, bristling with great mountains, wrinkled with the folds of compressed strata, and so thickly studded with volcanos that we listen readily to the assertion that it is pre-

eminently the great circle of dynamic action, and of volcanic as well as seismic energy. But if we examine these propositions in detail, if we project them upon a large-scale map, the less reason do we find to assign to this circle any real unity or continuity. On the contrary, it appears, with a little scrutiny, to resolve itself into a series of districts each of which, geologically and dynamically considered, is a law unto itself, with no more interdependence than several adjacent provinces in other parts of the world. It makes a radical difference whether we contemplate the distribution of these phenomena on a small-scale map or on a large one. Boussingault and Humboldt were much impressed with this fact in the Andes. The distribution of volcanos on such maps as then existed seemed to locate them in a single great chain. They found, on the contrary, scores of chains and *massifs*, sometimes crowded into narrow belts with correspondingly narrow intervales, sometimes spread out over broad belts hundreds of miles in width. So, too, on the small-scale maps the earthquakes seem to be near the volcanos, as if closely associated with them. They found volcanic vents and seismic epicentres separated by journeys of weeks with no discernible interrelation.

The questionable nature of the dependence so often asserted of earthquakes upon volcanic action has suggested the desirableness of detailed study of the distribution of the two classes of phenomena, and this study has been made with important results. It was begun in Japan by Prof. John Milne, assisted by Mr. Omori, and with the cordial support of the Imperial Meteorological Department. A system of postal-card reports of earthquakes from a number

of stations throughout the Empire was begun in 1880. As the results soon proved to be highly interesting, the number of reporting stations was much increased in 1884, amounting in 1885 to 968, and among them eight seismographic stations. Between 1885 and 1892 more than 8300 shocks were reported and recorded, and for each of them a separate map was drawn, showing the probable epicentre and the area sensibly affected by the shock so far as it could be inferred. These reports were then catalogued in the order of occurrence, a number being assigned to each shock. The list sets forth the date, the time (to minutes and seconds if possible), the area sensibly shaken, and the relative proportion of its conjugate axes. In conjunction with the catalogue a map of the empire was constructed, divided into squares of ten geographical miles on each side, every square having its own number. On this map the inferred positions of the epicentres were noted by dots.

This catalogue of Milne has certain advantages over others. It relates, indeed, only to a single seismic region or province, but gives, with as near an approach to completeness as is attainable, the records of all earthquakes occurring during a period of eight years. Within its field it renders possible a wider range of comparison with other phenomena (volcanic, lunar, solar, meteoric, etc.) than the larger catalogues of Perrey, Mallet, Fuchs, and De Ballore.

Among the conclusions which Milne derives from the study of his catalogue and map there are two which it is important for us to note here. The first is that "the central portions of Japan where there are a considerable number of active volcanos are singularly free from earthquakes. The

greater number of disturbances originate along the eastern coast of the empire and many of these have a submarine origin." "Lines 120 geographical miles in extent running in an easterly or south-easterly direction from the highlands of Japan into the Pacific Ocean, like similar lines drawn from the Andes westwards into the same ocean, have a slope of 1 in 20 to 1 in 30, and in both of these districts earthquakes are frequent. On the contrary, along the faces of flexures which are comparatively gentle, being less than half these amounts, which may be seen along the borders of most of the continents and islands of the world, earthquakes are comparatively rare. The inference from this is that where there is the greatest bending it is there that sudden yielding is most frequent."[1]

The other inference which Milne derived from this investigation was that earthquakes are frequent in those districts where there are evidences of secular elevation or depression still in process. This coincides with the deduction made many years before by Charles Darwin and Humboldt, though the evidence in support of it thus produced by Milne very greatly strengthens the thesis. Darwin's proposition is hardly more than an interesting but hazardous conjecture. Milne's only lacks an array of evidence supporting the reality of the elevations and depressions as clear as that which he furnishes of the precise distribution of the earthquakes to convert the conjectures into a demonstrated theorem.

Simultaneous with these investigations of Milne were those of Capt. De Montessus de Ballore along identical lines. He had for some years been gathering large quantities of

[1] *Seismological Journal of Japan*, 1895, p. xv.

statistics of earthquakes and had tested Perrey's laws in the light of extended data gathered with more scientific accuracy than was possible in Perrey's time. The conclusions he reaches are in general in harmony with those of Milne. He makes comparisons between the seismicity,[1] as he terms it, and the vulcanicity of 348 districts more or less disturbed by seismic and volcanic action, in which 9700 quakes and 5000 eruptions have been recorded. He draws a considerable number of conclusions from his comparisons, among which are the following:

1. In a group of adjacent seismic regions the most unstable (*i. e.*, most affected by quakes) are those which present the greatest differences of topographic relief.

2. The unstable regions are associated with the great lines of corrugation of the terrestrial crust.

These propositions, he observes, are subject to the qualification that seismicity is not strictly proportional to slopes, as the instability must necessarily be affected in a measure by the nature of the ground. It is probable, too, that other factors than relief are involved. In other words, the foregoing propositions express necessary conditions, but not all of them. But in any event the exceptions to the law of relief will not exceed ten per cent.

3. Rapidly deepening littorals, especially if they border important mountain ranges, are unstable, while gently sloping littorals are stable, especially if they are the continuations of flat or slightly accidented coastal plains. It

[1] The term *seismicite*, of which the above is a transliteration, is a very convenient word, and may be put into analogy with the term vulcanicity, which is sometimes met with. It means the degree to which a region or locality is disturbed by earthquakes.

has been possible in Chili, Peru, Japan, and the South Atlantic to mark out unstable regions whose epicentres are found not on land, but out at sea on submarine slopes of which the shores form the landward continuations.

4. These deductions are applicable to seismic regions and not to seismic centres individually. Though it is possible to indicate regions which present both volcanos and earthquakes there is no proof of interpendence between seismicity and vulcanicity in general. While there are earthquakes which are certainly of volcanic origin, the one phenomenon does not necessarily imply the other.

In the foregoing propositions of Milne and De Montessus de Ballore there may be danger that they will lead to misconceptions which the authors could not have intended to create, and which they would be among the first to correct if they appeared. In their earnestness to expose an error of older and widely accepted views they may have gone too far in the opposite direction. They have, indeed, furnished convincing evidence that earthquakes are not so universally or even so frequently associated with volcanic action as the world has been in the habit of believing. But the danger is that their discussions may create in some minds the inference that the two kinds of phenomena are wholly independent of each other. This would be a grave error and one that Milne and De Montessus de Ballore distinctly recognise. It still remains true that an important proportion of earthquakes are so associated with volcanos that we cannot reasonably doubt that volcanic action was the cause of them. But the investigations summarised above have satisfactorily shown that a much greater proportion give no

evidence of that origin. They indicate, rather, that they are the results of geotectonic forces and processes which we have yet to consider.

Those earthquakes which are unmistakably caused by volcanic action have certain characteristics which we may now proceed to compare with those which are so clearly the results of other causes.

Volcanic quakes, so far as we are sure of their origin, are seldom felt at great distances from their epicentres, even though their violence near the epicentre may be very great. The meaning of this, as will be more fully explained in a future chapter, is that the forces from which they originate are shallow. A few examples may illustrate this character.

A very striking instance is the earthquake at Casamicciola on the island of Ischia, on July 28, 1883, which was one of the most destructive to human life among the great quakes of modern times. Unluckily a large portion of the people at that favourite place of resort were assembled in the theatre, which suddenly collapsed and fell in a mass of ruins as if the Titan beneath had struck with his hammer an upward blow at the arch above his rocky prison with force enough to demolish everything standing upon the surface above. The town of Casamicciola was utterly wrecked, only one house being left standing, and the number of people killed by the falling ruins was nearly 1900. Yet in Naples, only twenty-two miles distant, the shock was noticed only by a few people as a faint tremor. The delicate instruments in Professor Palmieri's observatory on Mount Vesuvius showed no record of it, though the seismo-

graphs at Rome and Florence recorded some extremely light tremors.[1]

Casamicciola is situated upon the northern flank of Monte Epomeo, a volcano which may be regarded as practically extinct. Its last recorded eruption was in the year 1302, and none prior to that are known with certainty for more than eighteen hundred years, though some doubtful assertions of eruptions in the first two centuries of the Christian era have been put forward. The eruption of 1302 is quite authentic, and the large lava stream of that date is to-day very fresh in appearance and almost untouched by weathering or vegetation.[2]

A series of volcanic earthquakes on a much grander scale began on the southern flanks of Mauna Loa, on March 27, 1868. These shocks, at first light, continued to increase in force for six days. They came at intervals of only a few minutes, and every day there were many hundreds of them. On April 2nd they reached their greatest violence, and one shock in particular is described as being of the most terrible nature. The ground rolled in great waves, rapidly swaying in every conceivable direction, including the vertical. Stone houses and walls, chimneys and fragments of structures which prior shocks might have left standing were hurled down completely. Wooden houses were flung from their foundations. The rolling earth opened in great cracks on

[1] An excellent series of papers on this earthquake and the whole seismic and volcanic history of Ischia has been published by G. Mercalli in *Atti della Soc. Ital. di Sci. nat.*, vol. xxiv., and *Mem. del R. Instituto Lombardo*, Milano, 1884.

[2] Earthquakes had been frequent on the island of Ischia for centuries before the catastrophe of 1883. Casamicciola was destroyed by a powerful shock on Feb. 2, 1828, and Lyell, in his *Principles of Geology*, mentions seeing the houses in ruins the following year.

the crests of the waves, which closed together in the troughs. To stand was impossible either for beasts or men. Lying on the ground, it was at times necessary to keep the arms outspread to prevent being rolled over. The trees as the waves passed under them swayed violently, thrashing the ground and one another. At length, on April 7th, a radial crack, three-quarters of a mile long, extending up and down the mountain, opened about 5000 to 5500 feet above the sea on the south-western (and most gently sloping) flank of the mountain,[1] from which a sheet of lava shot up high into the air, sending a mighty deluge of fire to the sea. From this time on, the earthquakes rapidly died away, and a day or two later everything was quiet.

This eruption was of a highly exceptional character among the many historic outbreaks of Mauna Loa, which had been, so far back as their history extends, of a mild character, attended only with faint traces of seismic action, or even with none at all. The lava breaks out noiselessly, without tremors, and often the first that is known of it is the illumination of clouds when night comes, even though the mighty floods have been running many hours.

The feature of this quake with which we are here most concerned was its narrow localisation to the southern flanks of the island. The northern parts of the island were but slightly shaken. At Hilo, on the eastern flank, a single structure suffered material damage. A little farther north, in the deep valleys of the Hamakua coast and as far as Kohala, the shocks, says the *Hawaiian Gazette* of April 15, 1868, "though frequent, were comparatively light, except

[1] The slope of this flank is less than four degrees.

the severe one of Thursday, April 2nd, but even this, though causing some people to run out of their houses, did no damage to buildings. The Kohala plantation, chimney, and buildings were not injured." At Lahaina, on the island of Maui, about 100 miles distant from the epicentral points, the great shock of April 2nd was observed to shake pictures on the walls and rattle loose objects, while on Oahu, 150 miles away, the only persons who noticed it were a few in the second stories of buildings.[1]

This was one of the most forcible earthquakes which are assigned with certainty to a volcanic origin. Another almost equal to it was associated with the last historic eruption of Mount Ararat, on June 20, 1840. The shocks on this mountain appear to have been extremely forcible, all the villages near its base being shaken down in ruins, with a loss of life which was great relatively to the scanty population. But at Erivan, less than forty miles to the north-east, and a considerable city, the amount of damage was limited to a few cracked walls. At Kars, Erzroum, and Tiflis, all within 150 miles, the shocks were unnoticed.

The recent eruptions of Mount Pelée, in Martinique, and La Soufrière, in St. Vincent, gave rise to no far-reaching shocks or vibrations except some of a peculiar class to which we may here allude. It has sometimes been observed that during a volcanic eruption of the energetic kind, such as are exhibited frequently in the volcanos of the East Indian Archipelago, or in portions of the Andes, the

[1] Detailed accounts of this earthquake may be found in *Amer. Jour. Sci.*, series ii., vol. xlvi. The writer visited the locality in 1882 and questioned some of the witnesses of this affair who were then living and retained vivid recollections of it.

blowing off of the escaping gases at the vent, or the violent movement of the uprising lavas in the volcanic pipes, produce a sort of sustained chattering or vibrations of very short period, which are perceived both as a tremulous quiver and as a deep, humming sound at very great distances. It may continue for hours with long *crescendos* and *diminuendos*, while the vibrations of an earthquake, as ordinarily understood, last, usually, not more than one or two minutes. This phenomenon was quite marked in the recent eruption of the Windward Islands, and the sustained rumbling and chattering was recognised throughout the whole of that island chain. It has been repeatedly noticed in some of the old eruptions of Cotopaxi, and has been observed as far north as Cartagena and as far south as Lake Titicaca. The most remarkable instance of this kind appears to have been the eruption of Coseguina, on the Bay of Fonseca, Nicaragua, in 1835, the rumbling being heard at Kingston, Jamaica, at Caracas, Bogota, and Vera Cruz.

This class of tremors seems to be propagated with more than ordinary facility and to great distances through the deeper, more compact, and therefore more perfectly elastic, strata, and also through the water, and their sustained character tends to make them more easily recognisable than individual vibrations of the same, or even of somewhat greater power.

The eruption of Krakatoa in 1883, in which nearly one-half of the island and its volcanic mountain were blown up and scattered in small fragments over the Straits of Sunda, seems to have been about as energetic an occurrence as any of which we have detailed records. And yet at Batavia,

only 90 miles away, its vibrations were inconsiderable, and the most pronounced evidence of an earthquake was the sea-wave which entered the harbour. Similarly upon the adjoining Sumatran coast and upon the Preanger coast of Java the sea-wave was the principal phenomenon of a seismic nature, the vibrations alone causing no destruction.

The earthquakes which are associated with the eruptions of Ætna, though sometimes violent in the immediate vicinity of that mountain, seem never to have been strongly felt across the straits in Calabria. But the Calabrian quakes have been known to shake the Sicilian coast with great power. In the great Calabrian catastrophe of 1783 all Sicily was shaken, and in the city of Messina Sir William Hamilton[1] and Dolmieu[2] declare that not a house escaped injury, and most of them were partially or wholly destroyed.

Earthquakes are common around the base of Vesuvius and among the islands around Stromboli. But their tremors are seldom strongly felt in Sicily or on the Italian coast.

The restricted fields in which are manifested the vibrations of quakes resulting clearly from volcanic action may be explained by the comparatively small depth at which they originate and to the expenditure of a much smaller amount of total energy than is involved in the greater and more widely extended quakes.

It remains now to refer to the possibility that many quakes whose origin is unknown, or extremely doubtful, may, after all, be volcanic. This must be fully admitted,

[1] *Phil. Trans.*, 1783.

[2] "Dissertation on the Calabrian Earthquake" (translated), in Pinkerton's *Voyages and Travels*, vol. v.

and, indeed, it is in many cases highly probable. Evidences that volcanic action has taken place in the depths of the earth without visible, permanent results on the surface abound in ancient rock exposures. Formations of great geological age, once deeply buried and brought to daylight by secular denudation, show that lavas have penetrated surrounding rock-masses in many astonishing ways. Sometimes they have intruded between strata, lifting or floating up the overlying beds without any indication of escaping to the surface. Sometimes the lava breaks across a series of strata and finds its way into the partings between higher beds. Or it forces its way into a fissure to form a dyke which may never reach the surface. In one place a long arm or sheet of lava has in a most surprising and inexplicable manner thrust itself into the enveloping rock-mass, and in the older or metamorphic rocks these offshoots or apophyses cross each other in great numbers and form a tangled network of intrusive dykes. In other places the intruded lava has formed immense lenticular masses (laccolites), which have domed up the overlying strata into mountain-masses. These intrusions, almost infinitely varied in form and condition, are often, in fact usually, inexplicable as mechanical problems, but their reality is vouched for by the evidence of our senses. What concerns us here is the great energy which they suggest and their adequacy to generate in the rocks those sudden, elastic displacements which are the real initiatory impulses of an earthquake. They assure us that a great deal of volcanic action has transpired in past ages far under ground, which makes no other sign at the surface than those vibrations which we call an earthquake.

CHAPTER IV

DISLOCATION OR TECTONIC QUAKES

Chilian Dislocations and Associated Earthquakes—Similar Phenomenon in Cook's Straits—The Sonora Quake of May 3, 1887—The Dislocation which Caused it—The Great Mino-Owari Quake in Japan, October 28, 1891—The Great Dislocation of the Neo Valley—Its Destructiveness and Great Energy—The Bengal-Assam Quake of June 12, 1897—Probably the Most Formidable and Energetic of Record—The Characteristic Features of Tectonic Quakes—Their Vast Energy—The Immense Area throughout which they Are Sensible—Their Numerous After-Shocks—Omori's Study of After-Shocks—Scarcity of Preliminary or Warning Shocks—Tectonic Quakes without Visible Signs of Dislocation—The Andalusian Quake of December 25, 1884—The Charleston Quake of August 31, 1886—The Inyo or Owens Valley Quake of March 26, 1872—Absence of a Distinct Epicentre in Tectonic Quakes—Elongated Figure of the Meizoseismic Area—Agram Quake of November 9, 1880—The Middle Silesian Quake of June 11, 1885—Tectonic Origin of Great Sea Quakes—Schmidt's Study of Quakes in the Eastern Mediterranean

EARTHQUAKES which have been the accompaniments of dislocations of the strata have been the most impressive of which we have any accurate knowledge. Sometimes the dislocations are visible to the eye. Sometimes the sea covers them, but leaves them to be inferred from the vast sea-waves they are presumed to have caused. In some cases, with less confidence, with considerable uncertainty and hesitation, but not unreasonably, we may

conjecture their occurrence beneath the land, though no sign of it is revealed upon the surface.

The most remarkable, and perhaps the most fully described, events of this origin are those which have occurred in Chili. Among other observers they have received the study of Boussingault, Humboldt, and Charles Darwin, and their descriptions are among the classics of dynamical geology. On November 19, 1822, a great earthquake shook the Chilian coast for a distance of 1200 miles north and south, the greatest energy being shown about 100 miles north of Valparaiso. In the vicinity of that city the coast was found to have risen suddenly from 3 to 5 feet for a distance which has never been accurately ascertained, but which is known to have exceeded 35 miles. February 20, 1835, a similar event occurred at Concepcion, about 300 miles south of Valparaiso. Captain Fitz Roy, of the Royal Navy, who was making hydrographic surveys on that coast at the time, found that the land in the neighbourhood of Concepcion had been raised between 4 and 5 feet, and that for a distance of more than 130 miles south of Concepcion the elevation had been equally great or greater. Once more, on November 7, 1837, the town of Valdivia, about 200 miles south of Concepcion, was destroyed, and the coast in the vicinity was found to have been raised, for a great but unknown distance, from 5 to 8 feet.[1]

These upheavals, generating great and far-reaching earthquakes, may in a certain sense be regarded as types, and in a way they are not rare. Charles Darwin, in his narration, *Voyage of "The Beagle,"* says: "I have convincing proofs

[1] Darwin, *Voyage of the Beagle.*

that this part of the continent of South America has been elevated near the coast at least from four hundred to five hundred feet, and in some parts from one thousand to thirteen hundred feet, since the epoch of living shells." He finds his evidence in the raised beaches near the coast in which these shells abound and sometimes constitute almost the entire mass of the beach deposits. That this uplift has been going on by small and sudden movements, from a foot or two to ten feet at each shock, for more than two centuries, is attested by good evidence. The coast in many places is proven to be from twenty to thirty feet higher to-day than it was in the middle of the seventeenth century.

Sir Charles Lyell, in his *Principles of Geology*, gives in some detail a most interesting account of the sudden upheaval of a portion of a mountain range, with the accompaniment of a great earthquake, near Wellington at Cook's Straits,[1] in New Zealand, in January, 1855. Both the north and south islands of that colony have been affected by vertical movements during the nineteenth century, and these movements have been attended by powerful and far-reaching earthquakes. The changes wrought by these movements on the shores and adjacent topography have during that century been remarkable.

One of the most interesting instances of a dislocation generating a great earthquake was in the Mexican State of Sonora, near the Arizona boundary, on May 3, 1887.

This quake occurred in that plexus of mountains which

[1] The powerful earthquakes in the vicinity of Cook's Straits in 1848 are the subject of an interesting article in the *Westminster Review* of July, 1849.

are characteristic of the Sierra Madre of northern Mexico. The scene of the greatest disturbance was at first supposed to be in the valley of the Yaqui River,[1] about thirty miles south of the Arizona boundary. But it ultimately proved to be east of that river and farther south in a lateral range close to the Sierra Madre, known locally as the Sierra Teras.

The most striking feature was the long fault which had been produced, and which for a distance of over forty miles was plainly visible. The first five or six miles of its course was on the eastern side of the San Bernardino Valley, and it was not practicable there to decide whether the fault was a profound shearing of the massive rocks or a long landslip of the alluvial formations on the side of the valley. But farther southward the dislocation ascended higher up on the mountain slopes, cutting across rock spurs and cañons in a way that could not be mistaken and so conspicuous that for a course of several miles it was shown distinctly in the photographs. The displacement, as far as it was traced, varied from zero at the northern end to eleven feet about twenty miles farther south. Beyond that it diminished to

[1] This affair occurred while I was still engaged upon the study and discussion of the Charleston earthquake and burdened with other duties of an exacting nature. I took measures at once, however, to collect all attainable data concerning it, and was fortunate enough to interest an acquaintance in it who resided in Tombstone, Arizona, Dr. G. E. Goodfellow. Although the task was quite novel to him, he entered upon it with zeal and energy, and about a year later sent me an interesting report upon it which was published in *Science*, New York, April 6, 1888. In the course of 1887 the inauguration of the Irrigation Surveys of the United States was placed in my charge, and the burden became so great that it was impossible to pursue further the study of earthquakes. A large amount of data, however, was collected from many sources, and it is not impossible that a coherent account of this highly interesting and instructive occurrence may yet be wrought out in scientific detail and published.—C. E. D.

zero again, though a distinct crack was traced farther on until lost in the rocky wilderness of the Sierra Madre.

As this occurrence was more than two hundred miles from the sea, there was no possible reference to sea level to enable us to determine whether the movement was an uplift of the mountain range on the east or a downthrow of the valley on the west. There were two indications, however, which seemed in favour of the mountain uplift, though neither of them was at all decisive. The first was that the Yaqui River, which, except in a brief space on the west of the fault, showed no sign of having its gradients affected by it; and a river is a very sensitive index of changes in the slope of its bed. Another fact which is suggestive was the occurrence of another fault, formed at the same time, on the opposite or eastern side of the Teras Range, with its throw in the opposite direction. In other words, the range seemed to have been uplifted several feet between faults on either flank.

This earthquake was felt as far south as Durango, in Mexico, as far east as Fort Davis, in Texas, as far north-east as Las Vegas and Santa Fé, New Mexico, as far north-west as Prescott, Arizona. The average radius of distinct sensibility was not far from four hundred miles. It also disturbed the needle of the magnetograph [1] of the Coast and Geodetic Survey at Los Angeles, in California, a distance of more than six hundred miles. It was felt with notable force all along the coast of the gulf, from the mouth of the Rio Fuerte to the mouth of the Colorado.

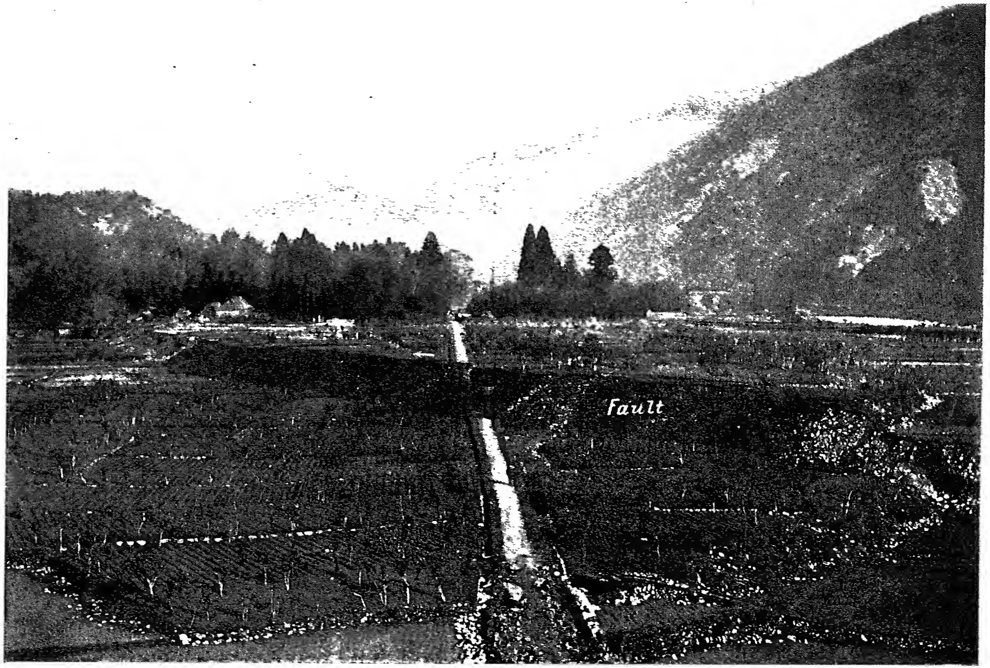
Another convulsion of the same nature and of greater

[1] The coincidence of time admits of no serious doubt on this point.

magnitude occurred in Japan on October 28, 1891, in the central and densely peopled provinces of Mino and Owari. At the head of the Ise Bay, on the southern coast of Nippon, is the city of Nagoya. Around it is the fertile delta plain, one thousand square kilometres in area, of alluvial soil brought down from the mountains to the northward, and supporting a population which in 1891 exceeded three hundred to the square kilometre. Twenty miles north of Nagoya is the provincial capital, Gifu, a town of about thirty thousand people. It is situated near the base of the mountains, which in a rather confused medley of ranges traverse lengthwise the great island of Nippon. Extending north-north-west from Gifu is Neo Valley, reaching far upwards into the mountain plexus, which is here composed of lower Paleozoic rocks and, near the main divide, of granite.

At the time of the earthquake a considerable fault was formed. Beginning at a point about fifteen miles east of Gifu it followed a direction varying from west-north-west to north-north-west across the great alluvial plain up the Neo Valley, across the main divide to the city of Fukui, near the coast of the Japan Sea, a total distance of about seventy miles. It thus crossed almost the entire breadth of Nippon. In the alluvial plain it might have been difficult to decide whether this was a true fault or merely a landslip of the soft, water-soaked alluvia. But along the mountain sides it was seen by Professor Koto, of the Imperial University, who devoted much time to its investigation, to cut the rocks overlooking the valley, while the slips of unconsolidated alluvia were abundantly displayed at the same time in the valleys below. The vertical displacement in

PLATE II.



The Fault whose Sudden Formation Caused the Great Mino-Owari Quake.

several places amounts to six metres, and three to three and a half metres were common,[1] and there was at the same time, says Professor Koto, a horizontal shifting of one and two-thirds to two metres in a north-westerly direction.

The earthquake was a very powerful one. The first and greatest shock brought down many thousand houses with their red-tiled roofs, killing and maiming multitudes of people. The official returns set forth 7279 killed, 17,393 wounded, 197,530 houses wholly destroyed, and 78,692 houses half destroyed. Extending northward from Nagoya to Gifu there was a nearly continuous street over twenty miles long where the houses were overthrown on both sides of the street, leaving only a narrow lane between two heaps of débris. The injuries to the railroad extending through these provinces, and especially the bridge crossing the Kiso River were remarkable, necessitating a new bridge and the reconstruction of about seventeen miles of very costly permanent way. But perhaps the most costly damage was inflicted upon the long dykes or levees (altogether more than five hundred kilometres in length) which hold the waters of the extensively ramified river system, passing through the great delta plain. Not far from Nayoga one of these levees was moved bodily over sixty feet back of its original position, carrying with it a thicket of bamboos and pines. The area sensibly shaken by this quake is estimated by Professor Koto at 243,000 square kilometres, or more than sixty per cent. of the Japanese Empire.

On June 12, 1897, there occurred in north-eastern Bengal

[1] B. Koto, in *Journal of Coll. Science*, Imp. Univ., Japan, vol. v., Pt. IV., 1893.

and western Assam an earthquake which Mr. R. D. Oldham, of the India Geological Survey, who investigated it, declared to be the most powerful and widespread quake of which history has given us any definite and detailed account, not excepting even the memorable event at Lisbon in 1755. The account of it has been published as one of the Memoirs of India Geological Survey, and so far as can be judged Mr. Oldham's claim seems to be justified. The epicentral area is of peculiar shape. Its southern boundary is a nearly straight line, about two hundred miles long, extending east-south-east; its northern boundary is a double sigmoid curve, and the included area is not far from six thousand square miles. Over the whole of that area the intensity was above No. X of the Rossi Forel scale, or greater than the intensity which prevailed in the city of Charleston, S. C., on August 31, 1886. Outside of those limits and within the further boundary where the shocks were distinctly sensible the area is computed at 1,750,000 square miles.

It was found that notable dislocations had occurred, and many of them both in vertical and in horizontal directions. Some of them were newly made faults, one of which was traced more than twelve miles with displacements measuring at several points over thirty feet. In other places changes of level athwart watercourses produced lakes, no less than thirty of these being noted, the largest being a mile and a half long and three-quarters of a mile wide. Their depths varied from one to over twenty feet.

Some of the triangles of the great Trigonometrical Survey were remeasured. While it is not possible to say how much absolute change has occurred, because the measured tri-

angles all lay in the epifocal tract, yet the relative changes were seen to be considerable. Relative changes in the heights of hills were found as great as twenty-four feet, and changes of twelve feet in their horizontal distances.

Mr. Oldham suggests that the true nature of the seat of the disturbance is a large thrust plane from which numerous faults branched off, some of which were plainly visible on the surface, while others were wholly subterranean, producing at the surface only gradual changes of level. In general the amount of displacement was extraordinarily large and distributed over a most exceptional extent of country. Within the disturbed tract the violence was excessive, especially in close proximity to the known fault planes. Not a masonry structure was left standing; monuments and tombstones were overthrown and broken, and even trees six or seven inches in diameter were snapped through their trunks by the swaying. All things considered, this occurrence may be regarded as a great type of the entire class of dislocation earthquakes.

The foregoing instances must suffice to illustrate the general character of those earthquakes which are plainly the results of dislocations of large masses of rocky strata. Though many others could be cited they are, on the whole, not common. It remains to advert to certain features which characterise them.

The first is the great extent of country throughout which they are felt, as well as their excessive violence and destructiveness near the dislocations which cause them. In this respect they far exceed the quakes of volcanic origin. The total energy expended in producing their vibrations must,

indeed, be enormous. Compared with it the vibrations caused by volcanic action involve a very much smaller amount of energy. How large a proportion of the total energy embraced in a great geologic downthrow or upthrow, or in a volcanic outbreak, is expended in producing elastic vibrations on the earth-mass we have not as yet the data for determining, though we may readily believe it to be a very small fraction only. This proportion may be greater in the dislocation than in the eruption. Whether it be because the total available energy is much greater or because the portion actually expended in causing vibrations is greater, the fact remains that dislocation quakes are the most potent and far-reaching, as well as the most destructive of any whose causes are known.

Another characteristic which distinguishes them from volcanic quakes is found in what are called after-shocks. All great dislocation quakes are invariably followed by repeated shocks, often of great force, though never equal to the primary one. The first few days or weeks after the catastrophe they are very numerous, keeping populations in a state of terror, and fearing a repetition of the awful visitation. But as time passes they become less and less frequent, as well as less forcible, and though they may continue for many years they gradually cease to be a source of apprehension.

These after-shocks have been the subject of a very interesting analysis by Professor Omori, of the Imperial University of Tokio.[1] He considers the after-shocks of the following important quakes which have been carefully and

[1] *Seismological Journal of Japan*, vol. iii., 1894, p. 71.

thoroughly recorded: the Kumamoto quake of 1889, the Mino-Owari quake of 1891, and the Kagashima quake of 1893. Within two years after the great quake of Mino-Owari—*i. e.*, from October 28, 1891, to October 31, 1893—3364 after-shocks were recorded at the Gifu Meteorological Station, of which 11 were "violent," 97 were "strong," 1809 were "weak," 1039 were "slight," and 408 were merely "sounds." Immediately succeeding the initial great shock strong shocks occurred very often, but these became rarer and rarer as the days advanced. Of the eleven "violent" shocks, ten occurred within the first four months, and the remaining one in September, 1892. The "strong" shocks all occurred within the first thirteen months, and the "weak" shocks all within the first twenty months.

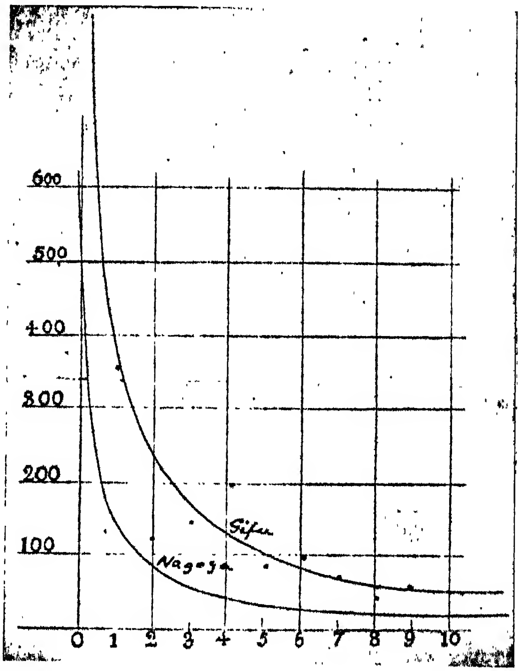


FIG. 5. Omori's Graphic Curves Showing the Number of After-Shocks of the Mino-Owari Quake in the Ten months Following the Catastrophe.

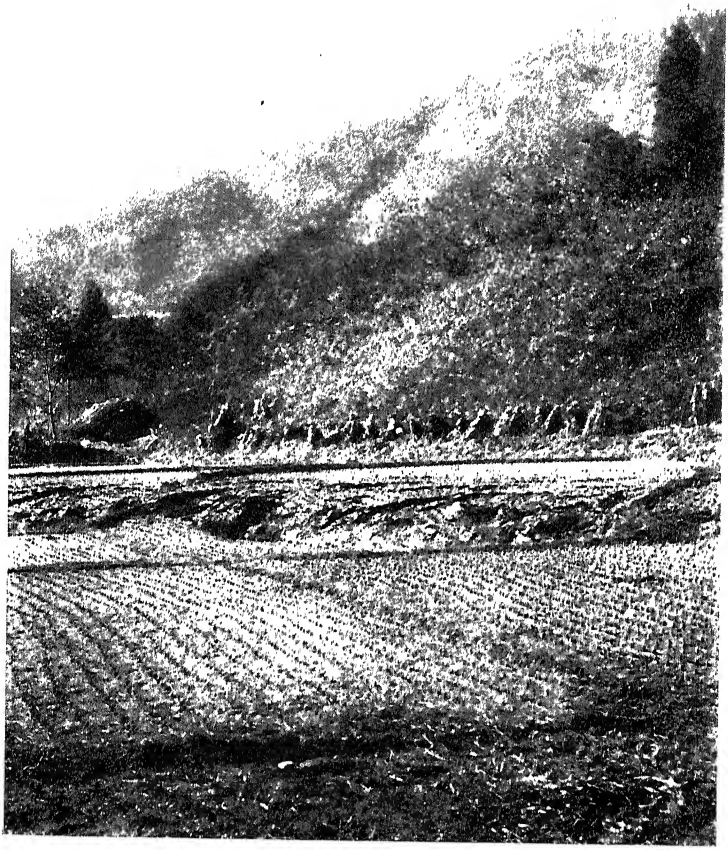
The total number of after-shocks of the Kumamoto quake recorded at Kumamoto to October 31, 1893, was 917, including two "violent" ones, which happened respectively July 28, 1889 (the same day as the great initial shock), and August 3, 1889, 72 "strong" shocks, 843 "weak" and "slight" shocks and "sounds."

Professor Omori makes inquiry whether there is a proportionality between the frequency of these after-shocks and the area affected by the principal quakes. This inference seems to be supported by the comparison between the Kumamoto and Kagashima quakes, but hardly so by the comparison between the Mino-Owari and the other two. And in general there seems to be no definite proportion, though in general it is everywhere true that the frequency of the after-shocks increases with the magnitude of the initial earthquake.

The usual assumption regarding these after-shocks is that the dislocated mass whose movement caused the original convulsion does not at once reach a position of equilibrium or stability, and that portions of the raised or sunken mass are subject to stresses which from time to time give rise to further movements until a final adjustment in a stable equilibrium is reached. Professor Omori suggests that if this failed to occur their absence might indicate "that the disturbed tract is probably being prevented from subsiding into a state of equilibrium and may be preparing for a second strong shock."

Volcanic earthquakes, on the other hand, are seldom followed by any considerable number of after-shocks, and in most cases are followed by none of any great magnitude.

PLATE III.



A Portion of the Mino-Owari Fault.

It seems as though the outbreak had relieved all tension, or exhausted temporarily the energy which might otherwise have produced them.

Both the volcanic and dislocation earthquakes are often preceded by what we may term preliminary shocks, though not always. They are much less numerous than those which always follow the great dislocation quakes and are seldom of great power. It may, perhaps, be regarded as most unfortunate that in neither of the two kinds of quakes are the preliminary warnings of such a character as to enable us to forecast what is coming, or even whether anything serious is impending. Sometimes great earthquakes, volcanic and dislocation alike, come nearly or quite unheralded. The eruption of Vesuvius, in A.D. 79, when Pompeii and Herculaneum were destroyed, gave some warning—enough to put many in terror and to induce them to leave the neighbourhood. But in 1794 one of the most powerful modern eruptions of Vesuvius was preceded by only a very few warning shocks.[1]

The fact that numerous after-shocks are highly characteristic of dislocation earthquakes has weight in suggesting the origin of some earthquakes whose causes would otherwise be extremely doubtful. Thus the quake which laid waste a large part of Andalusia in southern Spain, and which was investigated by a committee of the French Academy, may be with much probability attributed to a dislocation in the Sierra Tijada. While the committee saw no displacements except landslips and discovered no new faults cutting the

[1] A very eloquent description of this eruption was given by Leopold von Buch in *Beobachtungen auf Reisen*, Bd. ii.

rocks, the general character of their report is such as to suggest a movement of the Sierra. So, too, in the Charleston disaster, though no displacements other than those of the surface soil were indicated, the persistent after-shocks lasting through eight years or more are indicative of subterranean dislocation, and not of deep-seated volcanic action. The Owen's Valley earthquake of March 26, 1872, the most powerful one which is recorded in the history of our Pacific States, produced some extensive landslips, but its general features, including a large number of after-shocks, lasting through four or five years, suggest that dislocations not easily discoverable, and perhaps deeply buried, were the cause of it.

Another feature of the dislocation quakes is usually the absence of any strict epicentre. Instead of it the points of maximum intensity are arranged upon a long line, near the visible fault if there be any, or if not, then along a line, or elliptic space, which may be supposed to lie vertically over the line of dislocation deep in the earth beneath. Or, again, the places of maximum intensity on the surface may occur in an irregular, but always elongated, space overlying a subterranean focus of considerable extent and very irregular shape, any portion of which may give rise to powerful vibrations.

This elongated and somewhat indefinite figure of the epicentral tract, or district of greatest disturbance, is manifested in nearly, if not quite, all of this class of earthquakes, both when the dislocation is visible and when it is merely inferred. Examples of it, in connection with inferred dislocations not visible on the surface, are very numerous, and

we need to cite but few typical cases. One of the most notable was the Andalusian quake of December 25, 1884, where the epicentral tract, as determined by the French Commission, had a length east and west of over fifty kilometres and a breadth of about eighteen. So, also, in the earthquake at Agram[1] (Croatia) of November 9, 1880, the area of destructive action was an ellipse six (German) miles[2] long and three wide, extending north-east and south-west. Another quake which has been investigated with German thoroughness occurred June 11, 1895, in Silesia, near the towns of Strehlen, Reichau, Tepliwoda, and Peilau, about fifty kilometres south of Breslau.[3]

This quake exhibited a principal epicentral tract of decidedly elongated shape, its major axis lying north-east and south-west, with a secondary and still more elongated epicentral tract extending from the south end of the primary one a further distance of twenty kilometres westward. Two subterranean dislocations were inferred meeting at an angle of about 105°, and they appear to have been old faults which were increased suddenly,—the earthquake being an incidental result.

Finally, to cite only one more example, the Charleston

[1] An excellent and very detailed account of this quake, which was one of considerable force and destructiveness, though not of the first order, is given by F. Wahner in *Sitzungsber. der K. K. Akad. der Wiss.*, Bd. lxxxviii., 1883, pp. 303, etc. It is the subject of many references and examples by the seismologists of Germany and Austria, and was altogether a most suggestive incident.

[2] The German mile (Meile) may be reckoned roughly at four and a half English statute miles.

[3] "Das mittelschlesische Erdbeben vom 11, June, 1895," von Dr. R. Leonhard und Dr. W. Volz, Breslau, *Zeits. der Gesell. für Erdkunde*, Berlin, Bd. xxxi., 1896.

quake of August 31, 1886, exhibited, somewhat like the Silesian just cited, one well-marked epicentre with slightly elliptical isosemals, with a second one to the south of it more elongated, indicating that the subterranean source of the vibrations was not a compact locus, but an elongated one, giving off the most powerful impulses from loci near the ends of an elongated zone or vertical plane of disturbance.

It seems most reasonable to refer to the same cause those great events which have their origin beneath the waters of the ocean and which make themselves known to us by the great sea-waves they roll in upon neighbouring coasts. Sometimes they make themselves felt by the vibrations transmitted through the rocks. These vibrations may be powerful and destructive on land because they are near their origin, or they may have become enfeebled by distance from it. But in general their destructiveness is chiefly due to the vast waves rolled in upon towns or cities along the shores.

We seem to have no alternative but to believe that these mighty sea-waves originated in some displacement of large areas of the sea-bottom. No other origin is conceivable. Ordinarily it is a sudden *sinking* of the bottom, since the sea first withdraws from the coasts and then flows back, producing its most destructive effects. More rarely it is a sudden uprising at the bottom, and the incoming wave first appears. The great magnitude of these waves indicates a correspondingly great extent of sea-bottom involved in the displacement. To satisfy the requirements of such a wave as rolled in upon the South American coast at Arica in 1868 would require the sudden drop of many hundred square

miles of sea-bottom,[1]—perhaps of several thousand square miles, dependent upon the depth of water and the amount of the fall.

To this group of earthquakes belongs one of the greatest and most powerful convulsions of which modern history gives us any detailed and reliable account, the great event of November 1, 1755, which destroyed the city of Lisbon. It has been the subject of a large mass of literature, and has received a very interesting discussion from Lyell in his *Principles of Geology*. A valuable publication has recently been unearthed by M. Pereira and published in *Seismological Transactions of Japan*, vol. xii. Its value consists in the fact that it was written a few days after the catastrophe by an eye-witness of rather more than average understanding for the time, and contains no recital of some phenomena which in other accounts are such a heavy burden on the credulity—such as the sudden sinking of the quay on the riverside to unfathomable depths which later soundings have never been able to verify.

Much less known to English-speaking people are the great submarine earthquakes of the eastern Mediterranean described by Dr. Julius Schmidt.[2] Two of these, occurring October 12, 1856, and June 24, 1876, respectively, were of extraordinary extent. Schmidt at first was disposed to think that the epicentre of the first one (1856) was at the volcanic island of Thera, or Santorin, one of the Cyclades

[1] Enough is known of this great wave to make possible a rough computation of its length and amplitude in mid-Pacific, and, therefore, of the volume of water displaced. But the depth of the displacement and the amount of drop are, of course, quite conjectural.

[2] J. Schmidt, *Studien über Vulkan und Erdbeben*, Leipzig, 1884.

near the Grecian coast. But as data gradually accumulated he soon recognised that this occurrence could not be classed

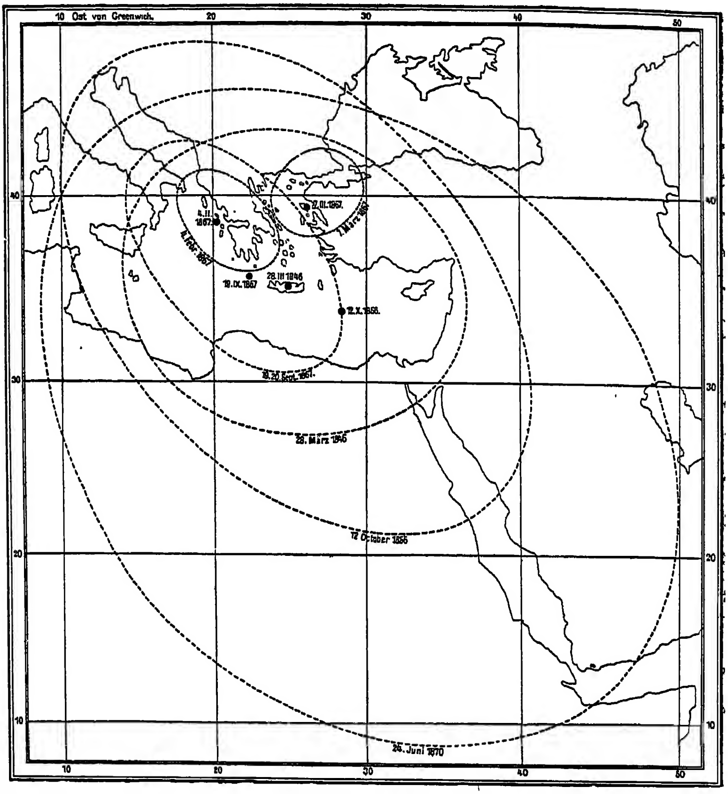


FIG. 6. J. Schmidt's Chart of Epicentres and Outer Limits of Sensibility of the Quakes of the Eastern Mediterranean from 1846 to 1870.

with volcanic quakes, whose foci were shallow and whose effects were not widely felt, and that it must have had a much deeper origin and involved a far higher order of total energy. His data finally led him to select for the epicentre

a point far to the south-eastward (about latitude 34°, east longitude 28° 30′), as the tremors had been felt as far north-westward as Venice, and in the opposite direction far up the Nile. The whole epicentral tract, therefore, was beneath the Mediterranean, very far from shore. No unusual motion of the water was reported, but as the quake occurred in the night a slight movement might not have been observed by any one likely to make it generally known.

The quake of 1870 appears to have been on a still grander scale. It was felt along the Red Sea coast of Arabia as far as Aden. In Cairo minor injuries were reported, and also in Alexandria, where the tremors were still more forcible. The whole Syrian coast was shaken with considerable force; Crete, Greece, Smyrna, and the islands of the Ægean felt the vibrations strongly. All southern Italy and all Sicily vibrated, and the rolling of the ground was generally noted in Naples, Messina, and Palermo.

This extraordinary quake is of special interest in several respects. Its epicentre Schmidt was quite unable to locate with confidence. No sea-wave was reported from any point on the coast. On the land there was no place where the intensity was so predominant over other localities as to justify its selection. Schmidt was then impelled to assign an extraordinary depth to the centrum. If this conclusion were correct it would make this earthquake an exception to all others whose depths we have any means of judging, and none of which furnish any reason for estimating their origins at greater depth than twenty miles and perhaps not more than fifteen. The paucity of Schmidt's data is such as to forbid any conclusions on this point.

CHAPTER V

SEISMOSCOPES AND SEISMOGRAPHS

Deceptive Nature of Observations without Instruments and Necessity for the Latter—Seismoscopes and their Limited Functions—Give Only an Instantaneous Phase of the Motion—Seismometers—A Steady-Point the Central Feature—Multiplying the Motion of the Tracers—Dr. Wagner's Device—Ewing's Duplex Pendulum Seismometer—Milne's Modification—Tests by Artificial Vibrations—Confusing Nature of Seismometer Traces—Necessity for Seismograph to Secure a Developed Trace—The Pendulum—A Long Period Necessary—The Conical or Horizontal Pendulum—The Gray-Milne Seismograph Described in Detail—Ewing's Circular Plate Seismograph—Dr. Agamennone's Horizontal Pendulum Seismograph Described in Detail.

THE first step in the experimental investigation of an earthquake is naturally to find some means of ascertaining as accurately as possible the real character of the movements which have taken place. The unaided senses are confused at the time, and the impressions they leave us, being unusual and very transitory, are only of the most general kind, and are surely misleading as to details. An instrument which can be set in motion by the quake and made to leave some automatic, intelligible record is needed. The earliest instruments designed for this purpose were of a very primitive character and were, moreover, based on a misconception of the motions of the ground during the earthquake. In most countries, especially those of northern

Europe and our own Atlantic States, such an occurrence is the event of a lifetime, or by a large proportion of mankind is not experienced at all. With most of the race knowledge of them is derived wholly from written accounts, and these, until seismology became a science, were vague and more or less imaginative. It is not surprising, therefore, that the notion prevailed that an earthquake consisted of a single "shock," or a very few shocks, instead of a sustained series of many hundreds of vibrations, as we now know. Conformably to this notion the earliest instruments were designed merely to record the fact that an earthquake had passed and the direction in which its impulse or "shock" had acted; for that the shock had some determinate direction was not doubted. Instruments of this class are termed *seismoscopes*. They do nothing more than announce the fact of an earthquake, and sometimes also the instantaneous direction of the force which set them in motion.

To devise something which will signal a decided tremor of the ground is the easiest kind of problem. Anything which trembles moves, and a very slight motion can be made to dislodge a delicately poised object, or to close an electric circuit so as to animate a magnet, or to start undulations in a dish containing a liquid. The number of such devices which have been experimented with or suggested is almost countless. But for the purposes and methods of modern research they are of slight value. Though of much historic interest they may engage but little attention here. Any cyclopædia, article "Earthquake," will give as much information about them as is really essential, and it is hardly worth while to repeat it.

The seismoscope, whatever be its mechanical construction, can give us only the result of an instantaneous phase of the movement of the earth, or of the support on which the instrument rests. In order to gain a more instructive idea of the movement we need an instrument which will trace the whole motion automatically from the beginning of the quake to the end, showing at every instant the direction, the amplitude, and the frequency, or "periods" of the vibrations. Such an instrument is termed a seismograph. There is an intermediate class of instruments which give continuous records of the movements of the ground and also show the directions and amplitudes, but they give them in such a complicated, tangled form that it is impossible to unravel them and show each vibration distinct from the others, as may be done with the record of the seismograph. These instruments are usually called seismometers.[1] Their results are very instructive and useful, and as the machinery is far simpler, less expensive, and more easily maintained than that of the seismograph, they are often very desirable. Their distinctive features will appear more clearly farther on.

The fundamental problem in all seismometers, including the seismograph, is to devise some object or mass which will remain at rest while everything around it and the very support which upholds it is in a constant state of vibratory motion. Such a mass is called a *steady-point*. To attain it is far from easy. But if it can be attained then we can conceive of a plate attached to it with a prepared surface capable

[1] The name seismometer seems to be generic rather than specific, the seismograph being a species of seismometer, and the generic name being applied specifically also to the less complex instruments.

of receiving a trace, and we can conceive of a tracer with one end attached to supports which vibrate with the earth, and the other end resting lightly upon the surfaced plate. Or, vice versa, we can conceive of the tracer attached to the steady-point and the surfaced plate attached to the vibrating supports. In either case a trace will be drawn on the plate showing the motion of the support relative to the steady-point. In the search for a steady-point, the first contrivance to suggest itself was naturally the simple pendulum at rest — a heavy bob suspended by a light and flexible wire. Though the point of suspension might oscillate it was thought that it would not be liable to impart any visible oscillation to the suspended weight unless the vibrations of the point of suspension were slow and of considerable amplitude. But experience quickly developed two difficulties. If the vibrations were slow and wide they invariably set the pendulum to swinging, so that it ceased to be a steady-point. If they were rapid and short the trace became a mere tangle and a small spot where its path was obliterated by crossing and recrossing itself. For the more powerful earthquakes, therefore, it proved at first to be of much less value than was anticipated. But for the less powerful ones it was conceived that it might be improved if it were practicable to multiply the motion of the tracer so as to give a much larger diagram of the relative motion.

As the use of the pendulum for a steady-point constituted the first step towards the attainment of a seismometer, so the enlargement of the trace may be considered as the second step. The first attempt to obtain an enlarged trace seems to have been in an instrument designed by Dr. G.

Wagner. In this device, Fig. 7, the tracer p is attached to the ring r r. To the upper side of the ring two small sockets are attached, one opening upward, the other downward. The upper socket receives a small knob projecting from the bottom of the pendulum W. BO is a light arm rigidly attached to the framework and ground and vibrating with it. Its outer end is turned up at a right angle and terminates in a knob entering the lower inverted socket. Thus two ball-and-socket joints are formed at O and S. When the earth vibrates BO vibrates with it. Assuming W to remain as a steady-point, the ring and tracer constitute a lever of the third order, and the motion of the lower end of the tracer multiplies the motion of BO in the ratio of Sp to SO. An obvious objection to this contrivance is at once suggested, though Dr. Wagner appears to have thought otherwise. Any horizontal motion of BO must communicate some degree of push to the pendulum, even though W be very heavy and SOp very light, and the push will be greater the farther the centre of inertia of SOp is from S. The pendulum will, therefore, swing and cease to be a steady-point.

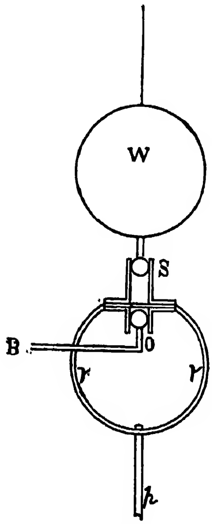


FIG. 7. Dr. Wagner's Multiplying Tracer.

Although this device was faulty, it contained the germ of something useful. It really multiplied the motion of the tracer, but sacrificed the still more essential requirement of a steady-point. To remedy the instability of the pendulum

and yet secure the multiplied trace, Professor Ewing proposed the "duplex" pendulum.[1] In Fig. 8 W is the main pendulum, consisting of a thick disc of lead with a hole passing through its centre, and suspended by wires attached to its periphery. W is a pendulum turned upside down, supported by a stiff rod. From the centre of its upper side a projecting knob enters the central hole in W, forming a universal joint. P is the tracer, having a knob on its lower end, entering the hole in W. Its upper end is recurved so that its point rests upon the tracing plate *t*. BO is the arm which actuates the tracer, embracing it by a ring at O, its other end being attached to the framework so as to vibrate with it. Now, imagine the earth to vibrate to the left, carrying with it the point of suspension of W. Then W will tend to swing to the left. But the same lurch will also carry *r*, the point of support of *w* to the left, and *w* will tend to fall *to the right*, and thus neutralise the tendency of W to swing to the left, provided the weights of the two pendulums are to each other as the lengths of their supports, and provided also the displacements of the two points of support are in the same direction and of equal amounts.

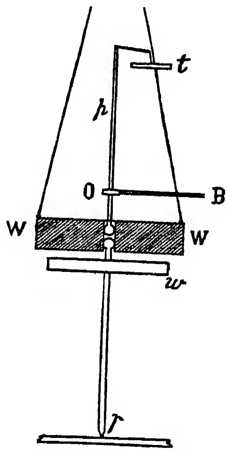


FIG. 8. Ewing's Duplex Pendulum.

This holds good also only through a very small arc of displacement. In practice the weight of pendulum W, divided by its length, must slightly exceed that of *w*. If it were

[1] *Loc. cit.*, p. 22.

less w would topple over entirely. With a slight preponderance to w it becomes a pendulum whose period of oscillation is very long, though the length of its suspension wires is comparatively short. The plate t in this device may be attached to the outer framework and vibrate with it, so that the trace becomes a record of the difference between the amount of the plate's motion and that of the tracing-point, and if the framework is quite rigid the direction will be the same in both. The two motions can then be disentangled from each other.

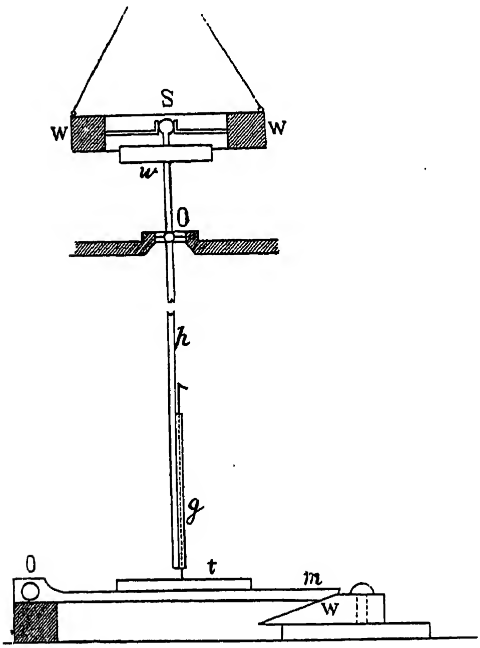


FIG. 9. Milne's First Improvement of the Duplex Pendulum.

A modification by Professor Milne of the foregoing design is indicated in Fig. 9. Here the inverted pendulum has a knob on its staff, which is supported at O upon gimbals, held by a part of the framework. The staff is prolonged through the knob, and this downward extension becomes

the tracer. At its lower end a very light tube of small calibre is attached, within which there slides freely a needle. The point of the needle rests upon a plate of smoked glass, as shown in Fig. 9. As the tracer oscillates around the point of support at O the slipping of the needle in the tube preserves continuous contact with the plate. The support upon which the plate *t* rests is constructed so as to admit of easy levelling by means of a wedge and set-screw. Here the plate vibrates with the earth and the trace represents the difference between the motion of the earth and the multiplied motion of the tracing-point.

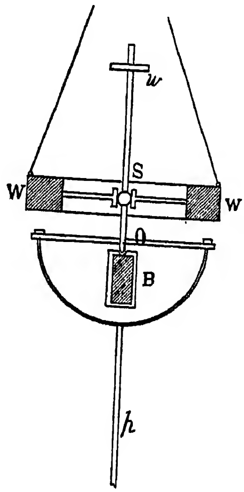


FIG. 10. Milne's Second Improvement.

Another modification of the duplex pendulum suggested by Milne is shown in Fig. 10, in which the secondary inverted pendulum *w* is above the primary pendulum W, instead of below. Here W is a heavy lead ring with a diaphragm across its interior. In the centre of the diaphragm is a hole, carrying a short cylinder open at both ends. B is a transom (in cross-section) attached to the framework, and the staff of *w* rests upon it, passing up through W and S, where it is enlarged into a knob or fillet which fits the short cylinder in the diaphragm of W. Near the bottom of the staff a cross-bar, O, is attached to it rigidly. To the ends of the cross-bar the branches of a semicircular band are attached. The

semicircle spans the transom. From the bottom of the semicircle the tracer *p* entends downward to the smoked-glass plate. In this form the weight of W is much less than in the preceding device, where *w* is below W, but the principle of the duplex pendulum is still preserved.

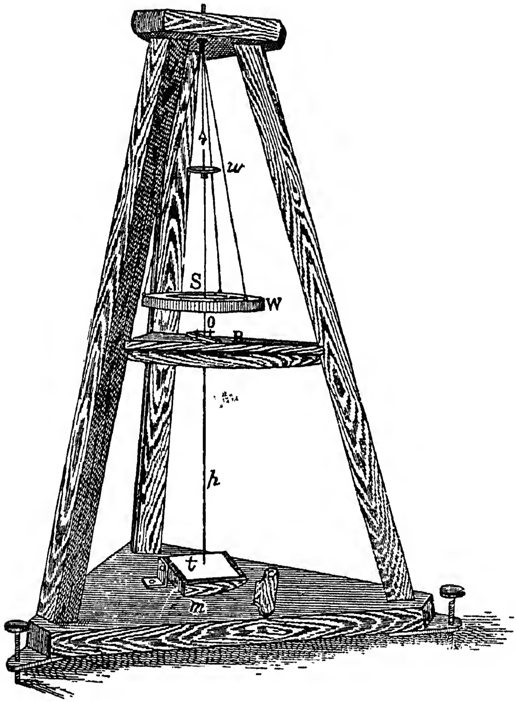


Fig. 11. Milne's Completed Form of Duplex Pendulum.

The several diagrams thus far explained are merely rough drawings of the most essential parts of experimental seismometers, and are employed to show as simply as possible the main principles involved. They are intended to secure a multiplied trace without disturbing the steady-point. The complete machine devised by Milne is shown in Fig. 11. The arrangement of the two pendulums is substantially the same in principle as exhibited and

explained in Fig. 10. It also shows in perspective the entire framework with its transom B, its base *m*, the cap *n*, carrying the suspension-point of the primary pendulum with means of adjusting it. The height of the tripod is two feet nine inches, and the side of the triangular base is two feet.

Enough has been said to explain the general character of the duplex pendulum seismometer and the principles upon which it operates, and it may now be well to inquire what measure of success attends its working. It is quite obvious that a shaking of the ground and whatever rests upon it must impart motion of some kind to the movable parts of the seismometer,—that the tracer will be moved with a multiplied motion and will trace a record of some sort upon a smoked-glass plate. But how can we be sure that the trace thus formed will correspond even with a rough approximation to the real motions of the earth? May not the path described by the tracing-point have a very different configuration from that described by a point on the ground? This question can be answered only by an experimental test in which the seismometer's trace can be compared with a trace produced by the same vibrating support upon a plate resting on the *quiet* earth. For instance, suppose the seismometer be placed upon a ramshackle table which can be easily shaken so as to simulate the motions of an earthquake. It is an easy matter to make the table record its motions with absolute certainty upon a plate resting upon the earth, which is now the best kind of steady-point. The seismometer can at the same time make its trace resulting from the shaking it gets from the table. The two diagrams can then be compared, and if they agree we may rely upon

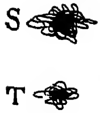


FIG. 12.

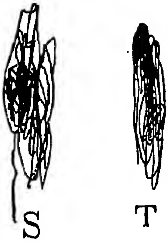


FIG. 13.

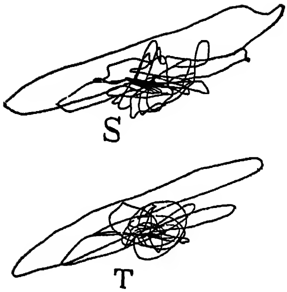


FIG. 14.

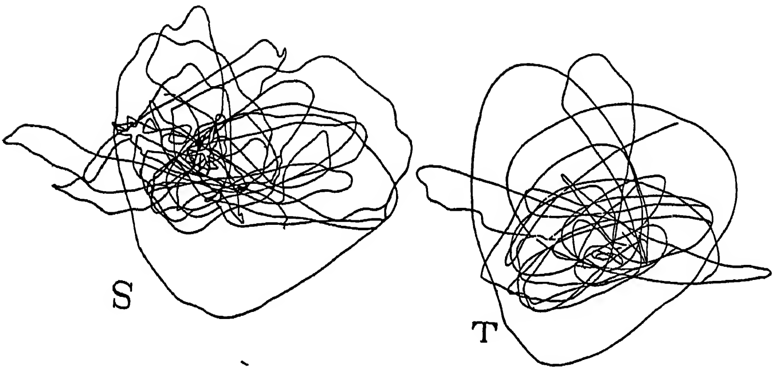


FIG. 15.

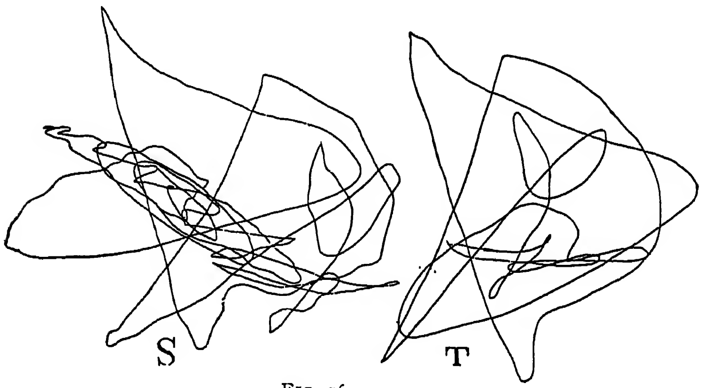


FIG. 16.

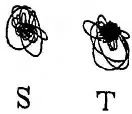


FIG. 17.

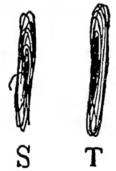


FIG. 18.

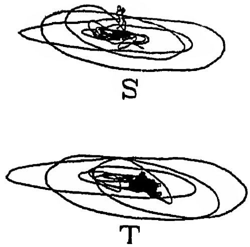


FIG. 19.

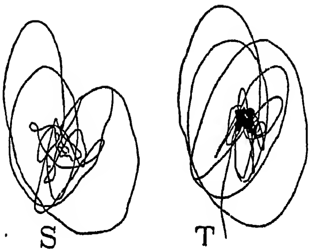


FIG. 20.

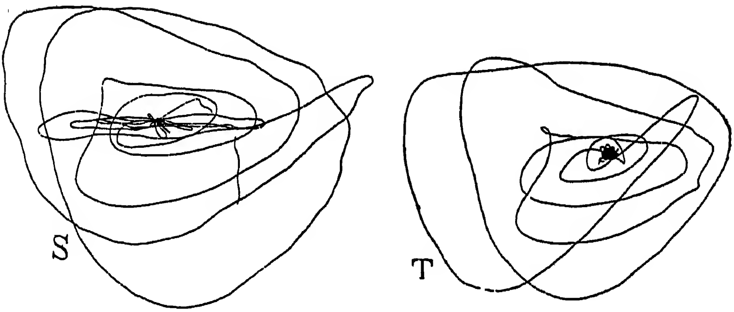


FIG. 21.

the veracity of our instrument. This experimental test has been thoroughly made in Japan, and Professor Milne has given an account of it,[1] the substance of which is as follows: Figs. 12 to 21 represent pairs of traces, one member of each pair denoted by T being the independent trace of the shaking table, the other member, denoted by S, being the trace made simultaneously by the seismometer standing upon the table. The first five diagrams, Figs. 12 to 16, represent comparisons of Professor Ewing's duplex pendulum seismometer, already described under Fig. 8. The second five, Figs. 17 to 21, represent similar comparisons of Professor Milne's instrument. It is seen that in each case when T and S are compared the traces are not quite identical, but show a good degree of approximation to agreement; indeed, a remarkable approximation, all things considered. The chief discrepancies in both of these instruments appear to be a tendency of the pendulum to acquire gradually a slight swing, which is kept up after the table is at rest, so that their traces show, near the close of the operation, wider amplitudes than the traces of the table and more of them. In the earlier and wider oscillations the agreement is very close.

On the whole, then, the duplex pendulum seismometer gives us a very fair notion of the movement of the ground during an earthquake, and we are justified in placing confidence in its record as an approximately truthful one,—so far as it goes. It enables us to measure the amplitudes of vibrations and the directions of the greater ones. But usually these vibrations are still so much tangled that we

[1] *Trans. Seismological Society of Japan*, vol. xii.

can seldom separate them. The requirements of thorough analysis call for some means of disentangling this knot, so that we can know not only amplitudes and directions in a few particular phases, but also the velocity of movement and the period of each oscillation, and thereby secure some data for estimating the force exerted, the duration of the entire quake, and the various changes in its phases. There is, however, an element of motion which we have not yet considered, and the "duplex" gives no account of it whatever. Yet it is a most important element and highly interesting and instructive. The investigator who wants to know all that is knowable must have a better instrument, giving much fuller and much more legible information. With this view the seismograph has been designed. But before attempting to describe it, it is necessary to examine some preliminary features. And first let us consider, briefly, some properties of the pendulum.

The period of a pendulum—*i. e.*, the time taken to perform a single complete oscillation—depends upon two quantities: first, the force which acts upon it, which in this case is the force of gravitation, and, second, the length of the pendulum—*i. e.*, the distance between the point of suspension and the pendulum's centre of gravity.[1] We can increase or diminish the period by lengthening or shortening it as we do in the case of a common clock. We can also increase the period by diminishing the *effective* force of gravitation, as is actually done in the duplex pendulum of Professor Ewing. For in this device two pendulums

[1] More properly, "centre of oscillation," though it happens to make no difference which term we use here.

oppose each other, and the tendency to swing—*i. e.*, the effective force acting upon the primary—is the difference between two opposing forces. And so the system as a whole has a greatly increased period of oscillation. By making the two opposing forces equal, all tendency to swing would vanish, and the period would become (in theory) infinite. There is, then, theoretically, no limit to the period which we can give to the duplex pendulum, no matter whether the length of the suspending wires be long or short.

Turning now to the simple pendulum, which was first resorted to in order to get a steady-point, we noted that it proved ineffectual and soon acquired a swing which vitiated its record. It is plain that the swing was caused by the lateral movement of the suspension-point, deflecting the suspending wire from the vertical. We noted, also, that the greater the displacement of the suspension-point the more quickly did the pendulum acquire its swing and the greater the amplitude of its excursions. We shall see farther on that earthquake vibrations of large amplitude usually have also long periods, while short amplitudes are usually of short periods. Hence large displacements of the suspension-point not only cause a greater deflection of the wire, but they give gravitation a longer time to act upon it. In any case, quite apart from amplitudes, it may now become apparent that the shorter the period of the earth-vibration and the longer the period of the pendulum, the less will be the tendency of the pendulum to acquire a swing, by reason of the displacement of its suspension-point. The periods of the earth-vibrations, of course, we

cannot alter, and we must take them as they come; but we can increase the pendulum period almost without limit, and that will serve the purpose equally well. The proposition then sifts itself down to this: The tendency of a pendulum to acquire a swing is in proportion to the ratio of earth-period to pendulum-period. The smaller we can make that ratio the less will the steady-point be affected by the vibrations. And the only way we can minimise it is to make the pendulum-period as long as is consistent with certain other requirements. In the duplex pendulum practical

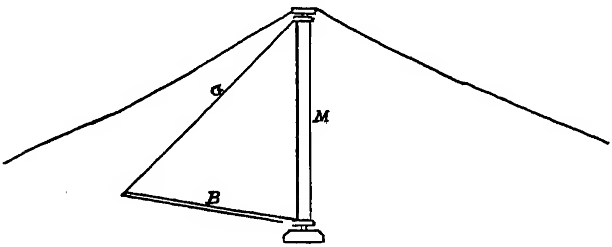


FIG. 22. The Conical Pendulum.

considerations limit the period which we can give it, and if we make it too long it will be liable to fail altogether. There is, however, a totally distinct form of pendulum to which we can give any length of period without practical difficulty. It is known as the *conical pendulum.*

We can see a conical pendulum almost any day. The derricks which are used in the erection of city buildings, the booms which a vessel lying at dock uses for unloading cargo, are examples of it. Fig. 22 represents roughly a common derrick, M the mast, B the boom, and G the stay. B revolves around the axis of M. If that axis is perfectly

vertical B will remain at rest in any position or azimuth in which it may be placed. But if the axis be slightly deflected from the vertical, B will seek the lowest possible position and will swing back and forth through it with a

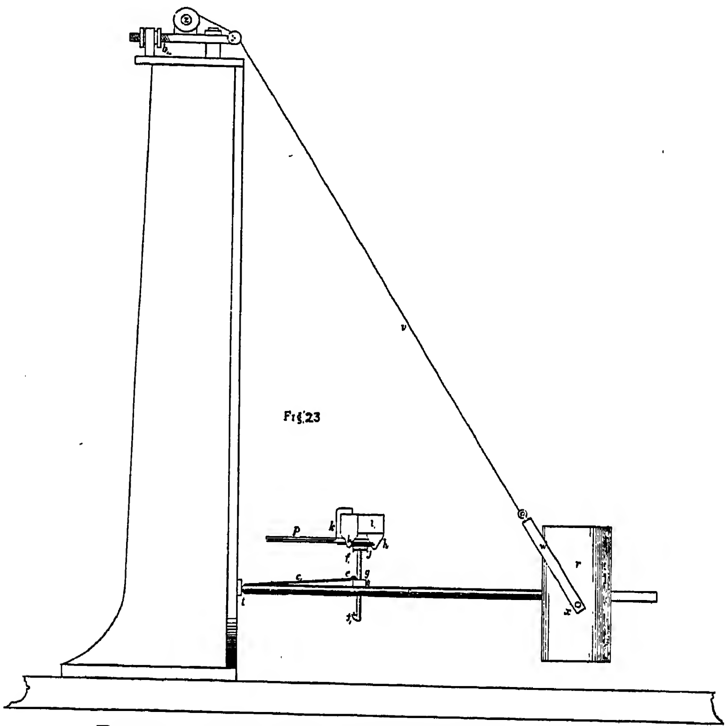


FIG. 23. Conical Pendulum of Seismograph. Elevation.

pendulous motion. B and G, as they swing, will each describe a part of a conical surface, and M will be the axis of the cones. The period of this oscillation depends upon the inclination of the mast. When it is quite vertical the period becomes infinite,—*i. e.*, there is no swing, and the

boom will remain in neutral equilibrium in any position. There is, then, no limit to the period which we can give to a conical pendulum, even though it be a small one, except the limits of accurate workmanship and adjustment. The use of this property in a seismograph was suggested by Professor Ewing (formerly of the University of Tokio), and it has proved to be a very valuable one. In each seismograph two of them are used, and they are hung in such

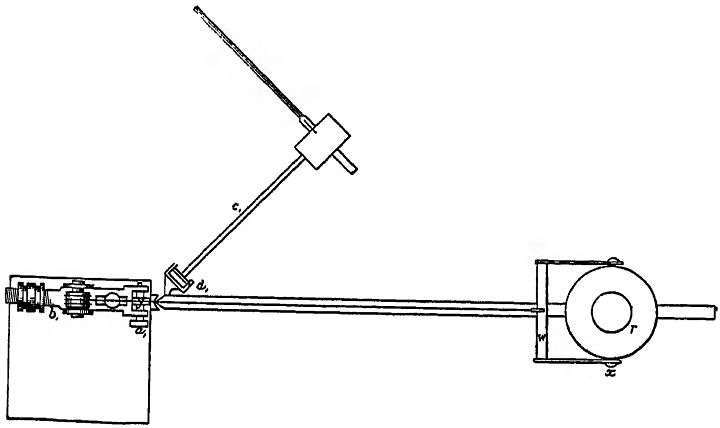


FIG. 24. Plan of Conical Pendulum.

a way that their booms (or struts, as they are called by Professor Gray) are at right angles to each other.

[Fig. 23 shows one of these pendulums in elevation and Fig. 24 shows it in plan, together with a contrivance for actuating the tracer. All other parts of the seismograph are for the present omitted from the figure. (The pendulum consists of a thin brass cylinder, *r*, filled with lead and held by a light tabular strut, *s*, furnished with a knife-edge at *t*, which rests against the bottom of a vertical V-groove

fixed to the supporting column *u*. The weight of the pendulum bob and strut is supported by a thin wire, *v*, attached in its lower end to a stirrup, *w*, pivoted at *x*, a little below and in front of the centre of gravity of *r*, and taken at the upper end over a small wheel, *y*, to a drum, *z*, around which the wire may be wound so as to adjust the level of the strut *s*. The position of the pivot, *x*, is so arranged that the knife-edge at *t* has little or no tendency to rise or fall, no matter at what part of the strut the cylinder *r* may be clamped. The wheel *y* is provided with adjusting screws, *a* and *b*, by means of which the top of the wire can be placed vertically above the knife-edge, or as much in front of or behind it as may be necessary to make the period of free vibration of the pendulum have any desired length.[1]]

Suppose now an earthquake-lurch moves the column in a horizontal direction perpendicular to *s*, then *r* will remain steady and *t* will move. Suppose the lurch to be parallel with *s*, then *r* will move with *u* and to the full amount of the lurch. Imagine next a second pendulum, R, hung with its strut at right angles to *s*, then the motion perpendicular to *s* which leaves *r* steady will move R, and the motion parallel to *s* which moves *r* will leave R steady. Suppose, again, that the lurch has an intermediate azimuth. The motion of *t* may then be resolved into two compartments, one parallel to *s*, and to that extent *r* will have a motion of translation equal to that component. The other component will be perpendicular to *s*, and the second pendulum, R, will have a corresponding translation, while the first pen-

[1] The description within brackets is taken from Professor Milne's paper in *Trans. Seis. Soc. Japan*, vol. xii.

dulum, *r*, will have no corresponding motion. Briefly, then, each pendulum is a steady point relatively to the components which are perpendicular to its strut, and by suitable mechanism can be made to record those components upon a moving ribbon of paper, or upon a revolving circular plate. Bear in mind that each pendulum traces only one component of the motion, and that one which is perpendicular to its own strut. The two components constitute the entire motion in a horizontal plane.

We come now to the measurement of the vertical components of earth-motion. This is also traced upon the paper ribbon simultaneously with horizontal components. The device for this purpose, employed in the Gray-Milne seismograph, is highly ingenious and is represented in Fig. 25, where *l* is a horizontal lever carrying at its free end the bob *m*. The other end carries transverse knife-edges bearing upwardly against a part of the framework *n* at *u*. The lever and bob are held up in a horizontal position by a large, bent spring, *oo*. Attached to the lever *l*, immediately over the knife-edge *u*, is a vertical arm of the lever at right angles to the latter. At the top of this arm is a second pair of knife-edges, *v*, bearing against a hook which is connected with a spiral spring, *t*, beneath. So long as the line of tension *t v* passes through the lower knife-edge *u* the pull of the spiral spring *t* has no result. But if the bob *m* be depressed, the knife-edge *u* becomes an axis of rotation for the lever *l*, and the arm *u v* rotates with it, carrying *v* to the left. The spring *t* then begins to act, tending to pull *v* still farther over. Meantime an increased tension is thrown upon the large spring *oo*, tending to pull the lever back to

its original position. Thus the spring *t* compensates in part the increased tension of *oo*, and the result is to give a longer period to the vertical oscillations of *l*. An ink-well,

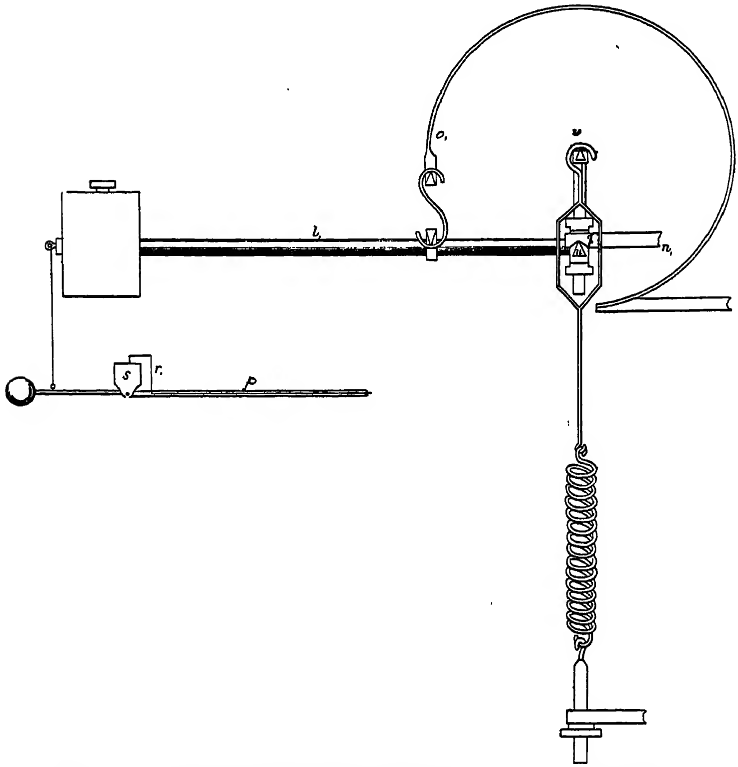


FIG. 25. Pendulum for Registering Vertical Component of Motion.

s, is attached to the framework and vibrates with it beneath *l*. To the lugs on its bottom is pivoted the tracing-lever, which is tubular and supplied with ink at the tracing end by the siphon *r*. At the other end of the tracing-lever is a knob and near it is a suspension-eye by which it is hung

to the free end of the pendulum lever *l*. When a vertical tremor comes the ink-well *s* is lifted, carrying the tracing end of *p* with it, with a multiplied motion, while *m* is a steady-point, and the weight of the knob keeps that end of the tracing-lever from rising when *s* is lifted. The tracer *p* will, therefore, move up and down relatively to *s* when vertical tremors come, and the tracer will write a sinuous trace upon a moving ribbon.

As regards the remaining parts of the seismograph, which are numerous and complex, it may be said that they are, in general, for the purpose of making the foregoing devices effective in writing out upon a ribbon of paper each of the three co-ordinates of the motion of the ground during an earthquake. They are mere mechanical contrivances of an ordinary kind, and of about the same grade of mechanism as a good watch or clock, and present nothing that is singular or distinctive. The really distinctive features of the seismograph are the pendulums and their mutual arrangement in such relations to each other that they separate the three components of the motions. The remaining parts and their assemblage may be described more briefly. Fig. 26 shows the Gray-Milne form of seismograph, with its two vertical pendulums, R and *r*, suspended in vertical planes at right angles to each other from the column U. The paper ribbon is unwound from the vertical drum, B, around the second drum, C, and is wound up on a third drum behind C and not visible in the figure. Three drums are used instead of two in order to get a uniform rate of speed for the ribbon, which would be difficult with two drums owing to the diminishing size of one roll of the paper and the

increasing size of the other. The train is driven by a falling weight or coiled spring, or any convenient sourçe of power. *ppp* and *iii* are the tracers and ink-wells, the lower tracer

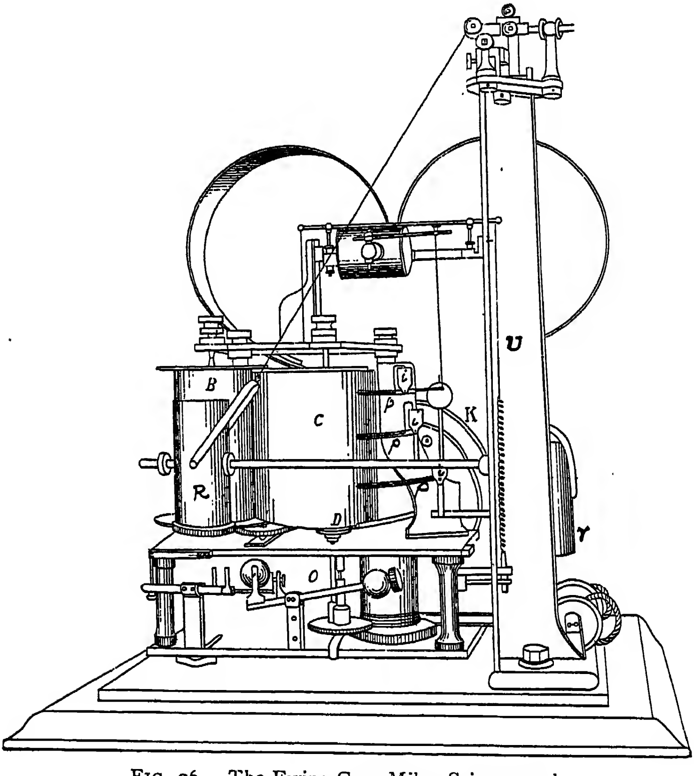


FIG. 26. The Ewing-Gray-Milne Seismograph.

being moved by the pendulum *r*, the middle one by the pendulum in full view, R, the upper one, by the pendulum for vertical co-ordinates. Underneath the table on which the drums rest is some mechanism for regulating their motion. When the earth is at rest the gearing is set so as

to give a very slow motion, feeding the ribbon say a quarter of an inch per minute. When a quake comes a delicately poised ball is dislodged and falls upon a lever. Instantly the gearing shifts so that the drums speed up and feed the ribbon forty or fifty inches per minute.

A good clock with second hands whose rate is known is attached to the instrument. At the falling of the ball and the tripping of the lever a movable dial is pushed up against the hands of the clock and withdrawn. The hands have each a projecting point smeared with printer's ink, leaving three dots on the dial which disclose the hour, minute, and second at which the quake began. Another device makes a mark at intervals of five seconds on the moving ribbon, so that the time corresponding to any particular co-ordinate of motion can be known to the fraction of a second by counting the marks back to the beginning.

Experience with the seismograph has been, on the whole, highly satisfactory. Comparative tests of the horizontal pendulums show that the traces they give undoubtedly represent faithfully the real co-ordinates of the motion of the ground on which the instrument stands, and this is true not only of the traces of horizontal motion, but of those of vertical motion as well. The conformity between the actual motion and the traces is very much closer than in the cases of the duplex pendulum seismometers already illustrated in Figs. 12 to 21. Examples of portions of these traces are shown in Fig. 27. They are actual earthquake records.[1]

[1] Accounts of these tests may be found with diagrams in *Memoirs of the Science Department*, University of Tokio, No. 9; in *Journal of the Science Department*, Imperial University of Japan, vol. i., and *Trans. Seism. Soc Japan*, vol. xvi.

Professor Ewing has designed a seismograph which traces the three co-ordinates upon a circular revolving plate instead of a moving ribbon of paper, and with excellent results. The only objection to it is that when the earthquake lasts longer than the time of one revolution of the plate the traces of the later portion of the quake overlie those of the earlier part, and when the duration is unusually long this superposition may be several times repeated, making it somewhat difficult to separate them. A diagram showing the traces yielded by the circular plate seismograph is given in Fig. 27, and reference will be made to it hereafter.

The composition of these three co-ordinates so as to give the resulting path of a point on the ground has been effected by the late Professor Sekiya, of the University of Tokio. The path is represented by a wire bent in such a manner as to correspond with it upon an enlarged scale. It is shown in three parts in Plate III, and is an admirable exhibition of the results of scientific research, reflecting much credit upon Professor Sekiya. The first of the three shows the path during the first twenty seconds of an earthquake occurring in Japan on January 15, 1887. The second shows the path from the twentieth to the fortieth second, and the third from the fortieth to the seventy-second second, at which time the vertical movement became inappreciable, though the horizontal vibrations continued considerably longer. The wire model was originally constructed upon a scale fifty times larger than the actual path, and the reductions of the engraver make the diagram about eight or ten times larger than the dimensions, the greatest

PLATE IV.

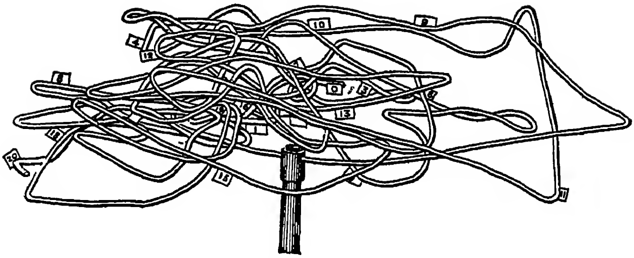


Fig.1.

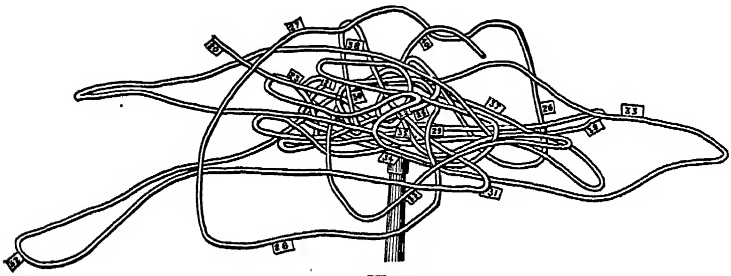


Fig.2.

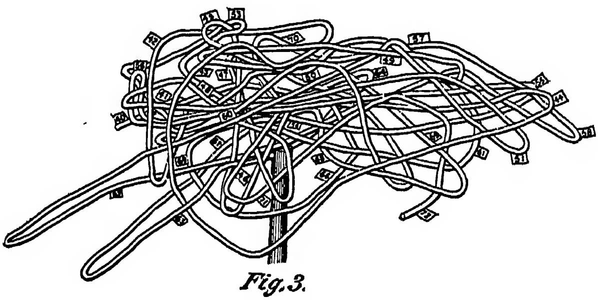


Fig.3.

Sekiya's Wires.

horizontal movement being 7.3 millimetres, the greatest vertical, 1.3 millimetres.

A highly perfected seismograph of the same class as the

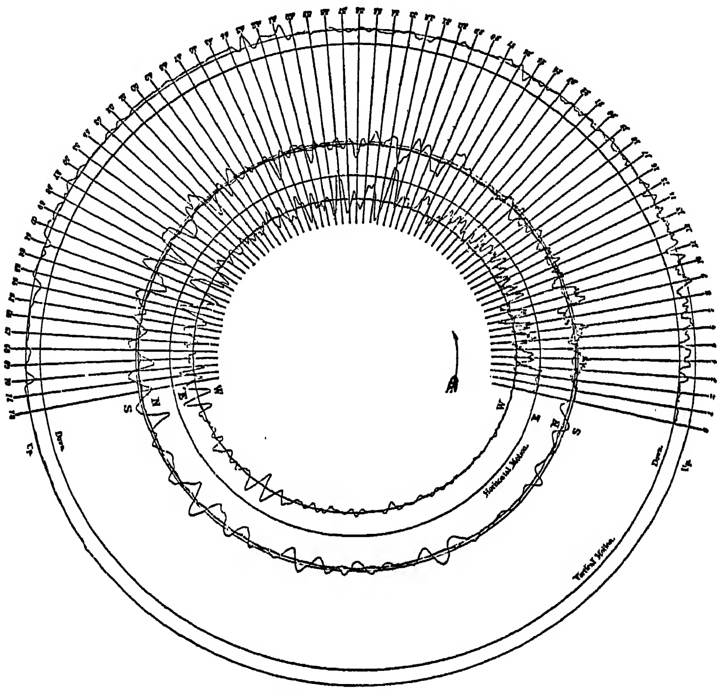


FIG. 27. Circular Plate Seismogram from an Ewing Seismograph.

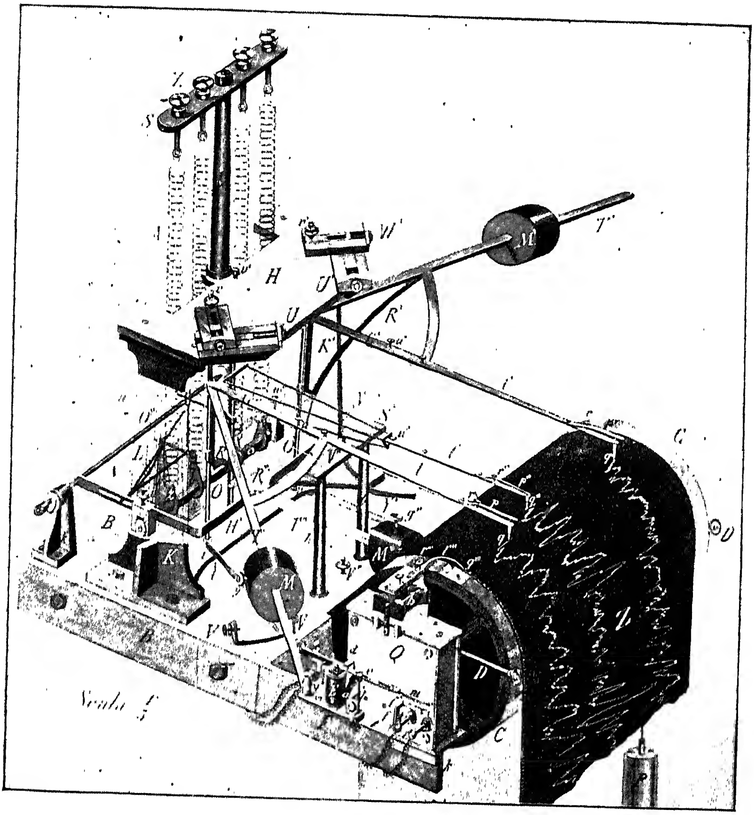
Gray-Milne has recently been designed by Dr. Agamennone, of the Central Bureau of Meteorology and Geodynamics at Rome. It embodies the developed results of about twenty years of the world's experience with seismographs, and is a very efficient instrument within its proper field. It is shown in Plate V, copied from *Bolletino della*

Societa Sismologica Italiana, vol. vii., 1901–02, and the accompanying description may here be condensed.

Upon the cast-iron base B are bolted two uprights, K,K′, carrying a platform, H, thus constituting a strong framework which supports two horizontal pendulums, MM′. The axes of the pendulums, OO′, are made of steel tubes. Near the top of each axis is attached a square bar, T and T′, on which bobs or masses, *MM′* (1½ kg.), can be made to slide towards or from the axes. These square bars or arms are upheld by the struts or brackets RR. Rigidly connected with each axis, at the same points of connection as the pendulum bars, is another arm, GG′. These are rigidly connected with the square bars also by metallic arcs of 45°. These arms, GG′, carry at their ends the tracing-stiles. The lower ends of the axes OO′ rest on conical steel points. Their upper ends also have steel points entering small conical cups bored in the ends of the milled screws *vv′*. The axes are adjusted with respect to a vertical line by means of the sliding carriage UU′, the upper slides of each pair carrying the screws *vv′*, which hold the upper ends of the axes.

Upon brackets FF, attached to the base B, is carried a drum, C, driven by a weight, P. The drum moves an endless band of smoked paper, Z. The stiles *ll′* are attached to the arms GG′ by means of "under-cuts" and are held in place by set-screws, *uu′*. Near the outer end of each slide is attached a flat spring, *pp′*, whose tension is controlled by the screw *rr′*. At the end of the spring is the needle *qq′*, whose point makes the trace. By means of the screw *rr′* the pressure and consequent friction of the needle upon the

PLATE. V.



Dr. Agamennone's Horizontal Pendulum Seismograph.

paper can be regulated, and any change in the inclination of the axes OO′ can also be accommodated at the needle-points.

To restrain any excessive swing of the pendulums a bar, YY′, is attached to the base of each axis, OO′. Its outer end has its swing limited between the stops VV.

The vertical component is obtained by means of a third horizontal pendulum, whose oscillation must be in a vertical plane, and whose axis must, therefore, be horizontal. *M*″ is the bob which slides upon a square bar. At R″ the bar splits into two branches, extending at right angles to it in opposite directions. Both branches are again bent at a right angle so as to become parallel and extend to the axis O″, which is a steel tube supported in bearings at the rear corners of the base-plate. A steel upright or column, K″K″, rises midway between these bearings and carries a cross-piece, S, at its top, in which are four milled screws, XX. Each screw has at its lower end a long helical spring. At the bottom of each helix is a ring through which passes a round bar, *o*, whose ends have bearings in jaw-shaped housings which slide upon the branches of the pendulum bar. The tension of the springs holds up the pendulum frame and the bob. (The pendulum frame is a rectangle composed of the two branches of the pendulum bar and the axis O″.) The milled screws, XX, permit the springs to be drawn up or lowered.

In the branch of the pendulum bar, which is in full view and near B, is a slot. Into this slot enters the end of a lever, L, which thus moves up or down with the pendulum frame. The other end of L is connected with an axis held

by two small uprights near L. At this axis the lever is bent vertically upwards at a right angle. The vertical arm is connected with a light bar attached to a vertical axis, *s* (between R and R″), moving it right or left as the lever L moves up or down. To the axis *s* is attached the arm of the tracing-stile. This arm consists of three very light brass tubes converging towards *u*″. At *u*″ the outer arm of the tracing-stile is attached, and is substantially identical in structure with the stiles already described.

A weight, P, is connected by a cord with the axis of the cylinder. The machine is at rest until a vibration starts a distant seismoscope, which closes a circuit and releases the weight P. This at once starts the cylinder and the paper band Z. At the near end of the drum is a train of wheels enclosed in the housing Q. They are driven by the axis of the drum and move the chronographic apparatus, which, in connection with circuits controlled by a clock, marks seconds on the edge of the paper band.

The apparatus is an excellent one, and, according to the reports of the Italian observers who have used it, gives remarkably good traces which show well vibrations whose periods are as short as the tenth of a second.

CHAPTER VI

ITALIAN VERTICAL PENDULUM SEISMOGRAPHS

Special Character of Italian Instruments—The Heavy Vertical Pendulum—Prof. Vicentini's Seismograph—Methods of Recording Vertical Components of Motion—Dr. Agamennone's Vertical Pendulum—Horizontal Pendulums of the German School—Dr. Rebeur Paschwitz's Pendulum—Milne's Horizontal Pendulum—Omori's Pendulum—The Bifilar Pendulum

THE movements which it is desirable to record by means of a seismograph differ widely in direction, amplitude, and period. Each kind of seismograph is adapted to recording a limited range of vibrations, or at least does much better work in some ranges and within certain limits than in others. In some ranges there are movements which some instruments do not record at all, though they may be highly efficient in the ranges to which they are adapted. Very few seismographs are capable of picking up tremors more rapid than ten vibrations per second. For vibrations whose periods exceed a minute, or even half a minute, and whose amplitudes are small, a very different class of instruments from those described in the preceding chapter is required. And it happens that these long, slow oscillations of very small amplitude are the ones which at present are the subjects of the greatest interest and of the most diligent inquiry.

The Ewing-Gray-Milne seismograph, just described, is the one with which English-speaking, and perhaps also Japanese, observers are most familiar. It is desirable, however, to describe the principal instruments used in Italy, which are very different in principle and have a wider capacity or range of performance, though in some respects less desirable.

What Professors Milne, Gray, Ewing, Sekiya, and Omori have done for seismometry in Japan has been accomplished along independent lines in Italy by Drs. Agamennone and Cancani, and Professors Vicentini and Grablowitz.[1] It is difficult to speak too highly of the valuable results obtained by the admirable service to which these gentlemen contribute, not only in devising instruments for investigation, but in all other branches of seismologic study.

The most important Italian instruments belong to the vertical-pendulum class. Their fundamental features consist (1) of a heavy bob of two hundred, four hundred, or even five hundred kilograms suspended by a wire to a bracket fixed in a wall, and having a length which may be as much as fifty feet (fifteen or sixteen metres); (2) they are provided with tracers which multiply the motion many times —fifty or even eighty fold; (3) the record is traced upon a

[1] The seismological researches of Italy are conducted in connection with the meteorological observations which are within the general administration of the Bureau of Meteorology and Geodynamics at Rome. This bureau has long been under the very efficient charge of Prof. Pietro Tacchini, and is a branch of the Department or Ministry of Agriculture, Industry, and Commerce. Dr. Agamennone is assistant in charge of the Geodynamic Division in the Central Bureau at Rome; Dr. Cancani has charge of the Geodynamic observatory at Rocca di Papa, about sixteen miles from Rome; Professor Vicentini is Director of the Physical Institute in the University of Padua; Professor Grablowitz has charge of the observatory at Casamicciola.

ribbon of smoked paper which moves fast enough to separate all tremors which the instrument can record. The Italian instruments of the vertical-pendulum class are the same in principle and differ only in minor details. It will suffice, therefore, to describe one of them, and the seismometrograph of Professor Vicentini is selected.

It consists essentially of a large leaden bob (50 kg.), Fig. 28, hung by three metallic rods united above to a brass cap hanging by a steel wire, 2 mm. gauge. Each of the three metallic suspension-rods is provided with a turnbuckle which can be used to level the bob, and thus make its axis vertical. The upper end of the suspension wire is attached to a screw which pierces a stout iron bracket fixed in the wall. By means of the screw the system can be raised or lowered a little, so as to hang just above two iron beams, SS, which are intended to catch the weight if the suspension wire breaks. Three screws, VV, held by an iron band surrounding, but clear of, the bob

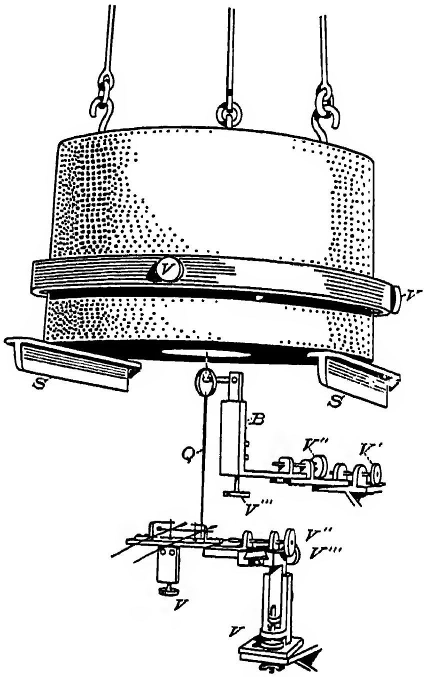


FIG. 28. Vicentini Vertical Pendulum.

prevent too great an oscillation. The length of the pendulum is about 1½ metres, and its period for a complete oscillation is 2.4 seconds.

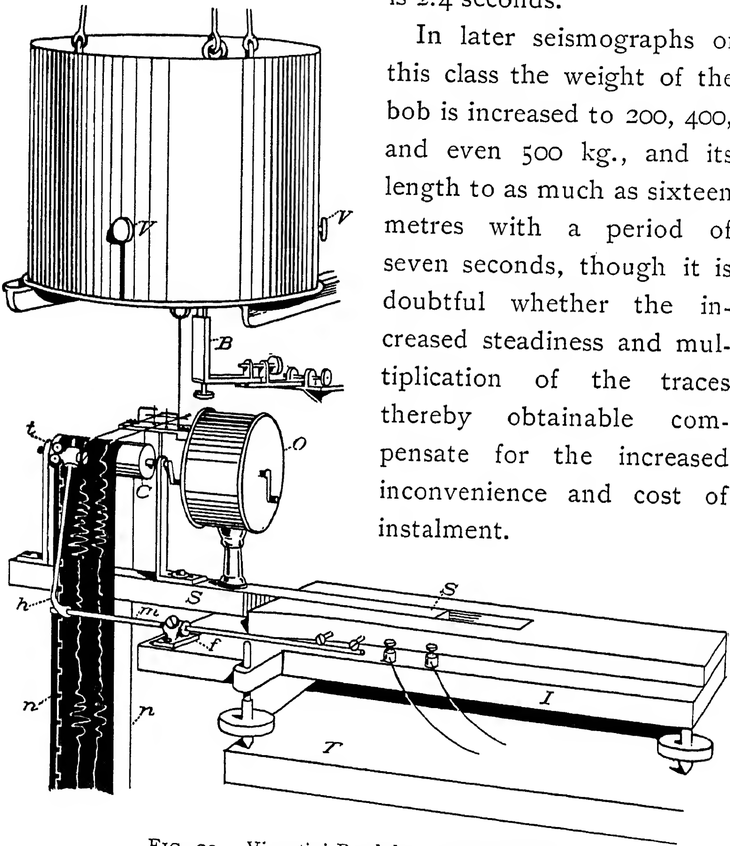


FIG. 29. Vicentini Pendulum and Recorder.

In later seismographs of this class the weight of the bob is increased to 200, 400, and even 500 kg., and its length to as much as sixteen metres with a period of seven seconds, though it is doubtful whether the increased steadiness and multiplication of the traces thereby obtainable compensate for the increased inconvenience and cost of instalment.

In the centre of the base of the bob (which is very slightly conical) is bored a small cavity, covered by a brass plate. A small, bevelled hole is bored in the plate so as to coincide with the vertical axis passing through the centre of gravity

of the bob. When the ground oscillates the bob acts as a steady-point,-- if the oscillations are not too prolonged. With slow vibrations — or with vibrations of a particular value—the bob acquires oscillations of its own. These vibrations, whether of the ground or of the pendulum, are made manifest through the multiplying lever *l*, of which an enlarged view is given in Fig. 30. It consists of a tube of thin aluminum foil, attached to a ring of the same metal, *m n r*. A steel needle, *q c*, pierces, and is soldered to, the ring at *m*. A second needle, *o p*, is attached to the tube *o n*. The point *c* of the upper needle rests in the conical cavity of a glass cup carried by the support B, which in turn is rigidly attached to the same wall which carries the pendulum. The support B is provided with adjusting screws, whereby its position beneath the bob can be so controlled that the upper end of the needle *q* can be brought vertically under the bevelled hole. In this position a few turns of the third screw *v″* raises the bearing which carries the glass cup, and with it the lever *l*, so that its upper end, *q*, enters the bevelled hole as far as desired, say *t*.

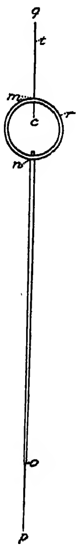


FIG. 30.

The bob being a steady-point, the aluminum lever becomes a lever of the third order with its fulcrum at *t*, with *t c* as its short arm and *t p* as its long arm. The *point d'appui* is at *c*. The multiplication at the lever-end *p* is proportional to the ratio between *t p* and *t c*. But if the bob oscillates, the point *c* becomes the fulcrum of so much of the movement of *p* as depends upon the swing of the bob, and in calculating the multiplication it is necessary to take account of it.

The recording device is so designed as to multiply again the motion of the lever-end *p* and to resolve its motion into two rectangular co-ordinates. The first multiplication of *p* is already about fivefold, and this is again multiplied ten-

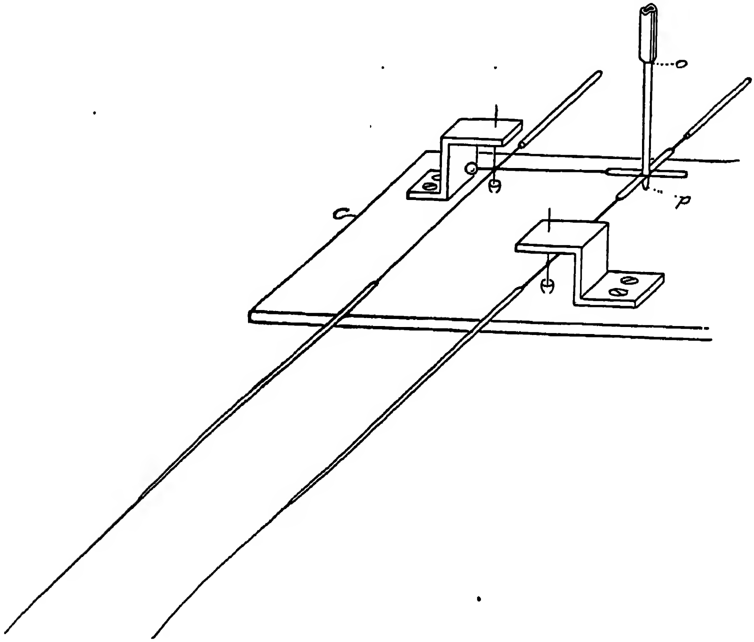


FIG. 31. Multiplying Tracers of the Vicentini Pendulum.

fold, or more, in the recording device,—*i. e.*, fiftyfold or more altogether.

The decomposition of the motion into two co-ordinates is effected by means of two aluminum tracing-stiles made as light as possible. They are carried on the support C, which is fixed to the wall. The stiles are shown in Fig. 31. One of them is straight and provided near its rear end with a

hole to receive a vertical axis. The rear part is attached to a small, slotted link of polished steel, the slot being just wide enough to admit easily the lower end, *p*, of the vertical multiplying lever, Fig. 30. Back of the steel slot is a very small counterweight so adjusted as to minimise the bearing weight and friction of the tracing-end upon the smoked paper. The other stile is bent at a right angle at the axis, and the bent arm is provided with a slotted link similar to that of the first stile and crossing it, a little below or above as may be convenient, at a right angle. The needle *op* also penetrates this slot. The counterweight of this stile is in the direct continuation of the stile back of the axis.

In Fig. 29 the foregoing parts are shown together in their proper relations. Upon a shelf, T, built into the wall, is a support, I, resting on three levelling screws. Upon it slides a carriage, SS, whose motion is controlled by guides. On the carriage stands a clock, O, whose axis is connected to the axis of the horizontal cylinder C. This cylinder, moved by the clock, revolves once per hour, and brings under the tracing-stiles an endless ribbon, *n*, of smooth smoked paper. The height of the cylinder C is carefully adjusted beneath the points of the stiles so as to receive the minimum friction.

This delicate adjustment of both the stile-points is not secured directly, since it is impracticable to set them so accurately. To meet this difficulty it is necessary to have recourse to the levelling screws of the table I, which are introduced for this purpose, and the adjustment is further equalised by raising or lowering the axes of the stiles by screws which control their vertical positions.

A good many kinds of clock might, perhaps, be used, but to ensure the proper feeding of the paper ribbon Professor Vicentini employs a good chronograph with a Depretz signal attachment, or ticker, connected with a pendulum clock which closes an electric circuit every minute. The signal is placed in front of the cylinder C at *t*, and can be raised or lowered on a sliding-rod, *n*, whenever the signal becomes indistinct. The signal also can be adjusted nearer to or farther from the edge of the paper by its attachment

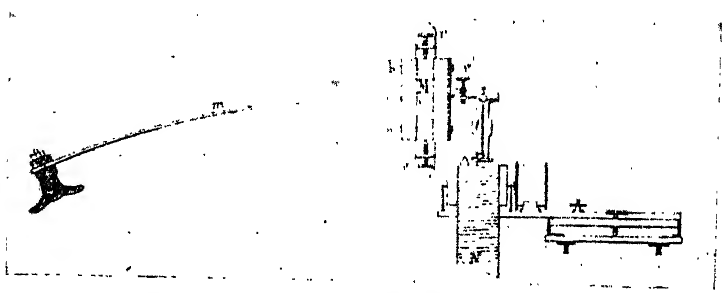


FIG. 32. Device for Vertical Motion, Vicentini.

to the horizontal rod *m*, which slides in a bearing fixed to the support I at *f*, and is clamped there by a set-screw. The signal makes jogs every minute upon the edge of the paper, and thus supplies the time scale. At every hour the signal is duplicated.

The pendulum bob is protected from air currents by a sheet of metal beneath, with a circular opening for the multiplying lever, and by a jacket of sheet metal around it through which the screw-heads VV project.

The paper ribbon moves at the rate of about 2 mm. per minute. The instrument is so sensitive that it is never at

rest. Both stiles are in a continual state of unrest, especially the one recording movement in a direction perpendicular to the wall.

The vertical motion is obtained by a separate device,

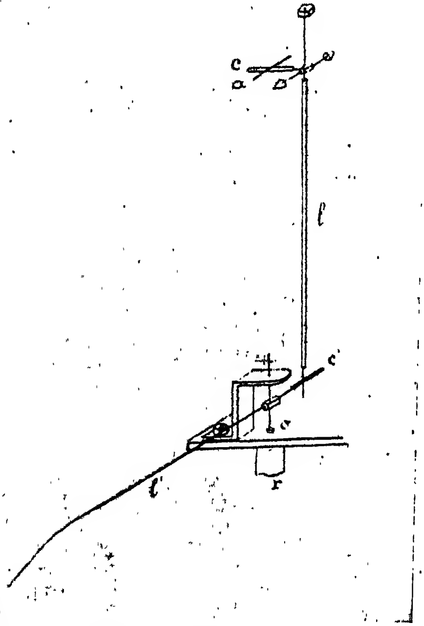


FIG. 33. Stile for Tracing Vertical Motion.

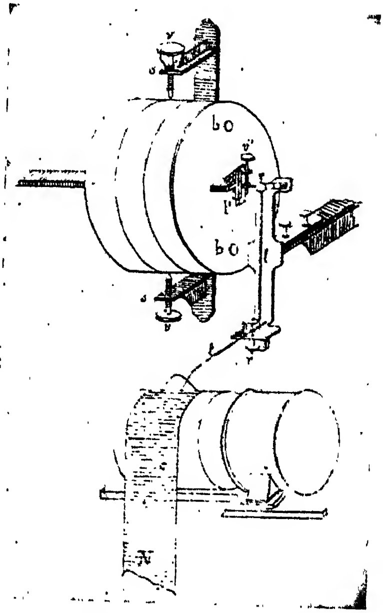


FIG. 34. Pendulum for Vertical Motion.

which has been added since the foregoing apparatus was first illustrated and published. In Fig. 32 *M* is a cylindrical mass of lead (45 kg.), at the end of a flat spring, *m*, 1.50 m. long, 75 mm. wide, and 10 mm. thick at the proximal end and 7 mm. thick at the distal end. The proximal end is firmly held by a pillow-block, S, sunk in the wall. The block

is inclined sufficiently to make the distal end horizontal under the sag of *M*. Two screws *vv* in a bracket attached to the wall limit the vibratory motion of *M*.

The multiplying device consists of the lever *l* and the stile *l'* (Figs. 33, 34). A slide, P, in an undercut guide-frame with a milled adjusting screw, *v''*, is attached to the extreme outer end of the flat spring *m*. A horizontal needle, *a*, controlled by *v''*, Fig. 35, enters the slotted arm *c*, Fig. 33. This arm *c*, united to the lever *l* at a right angle, constitutes a bent lever with its fulcrum at the angle of junction. The lower end of *l* terminating in a needle enters the slotted end of the stile *l'*. The remaining action is the same as for the horizontal components.

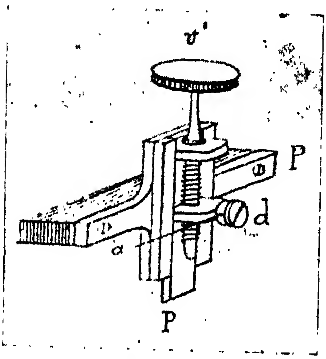


FIG. 35.

The vertical pendulum of Dr. Agamennone belongs to this class. Being designed to record the smallest as well as the largest movements of the ground, he calls it a micro-seismometrograph. In order to obtain steadiness of the pendulum with a very great multiplication of the traces he uses a very heavy bob (500 kg.), Fig. 36, in shape a very short cylinder 60 cm. in diameter. It is composed of four discs of lead, each divided into twelve sectors to facilitate assembling and dismounting. The forty-eight sectors are held in place by three iron discs assembled by three bolts and by dowels within the mass which prevent all displacement. The bob is suspended by three iron rods.

each provided with a turnbuckle, and uniting above in a cap attached by a single wire, as in the Vicentini pendulum. The whole length of the suspension is about ten metres, giving a period of 6½ seconds.

Beneath the bob is a strong cast-iron cross bolted to a masonry foundation, its four arms turned up at a right

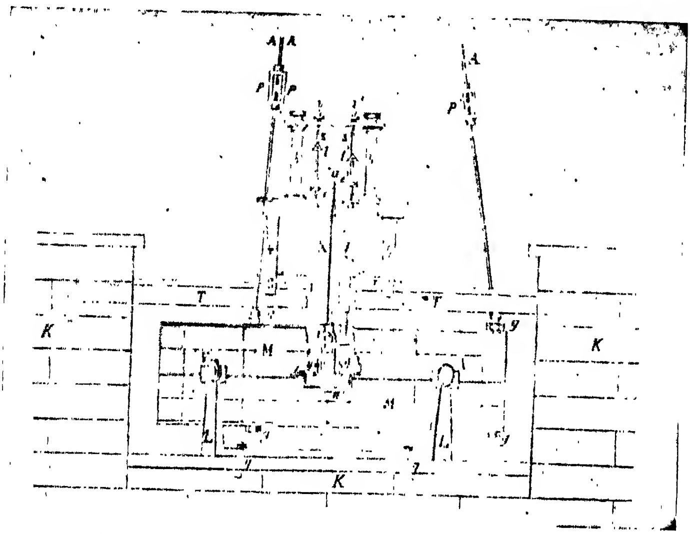


FIG. 36. Dr. Agamennone's Vertical Pendulum Seismograph.

angle to form the columns LL. Each has a screw, V, to restrain any excess of oscillation. Above the bob are two iron beams, TT, one behind the other, set in the masonry piers KK. They support two transoms which in turn support the tracing device.

A well is made in the bob deep enough to establish a pin, *a'*, at its centre of gravity, which actuates the vertical lever X. This lever, X, consists upwardly of a tube of aluminum

foil. To its lower end is attached a fork, in the axilla of which is a recess permitting it to rest securely upon a steel point, *i*, projecting from the lower end of the support U. This support, being attached in turn to the beam, vibrates with the ground or pier, communicating its motion to the

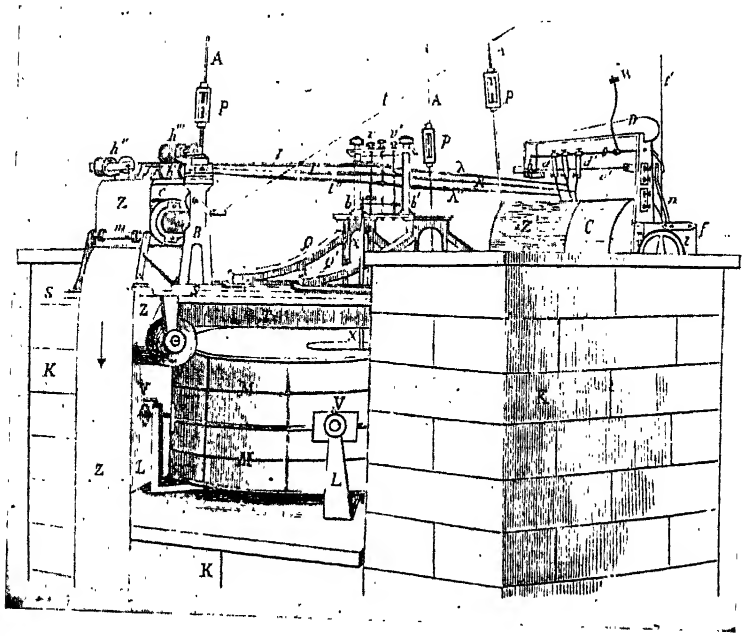


FIG. 37. Tracing-Stiles of Agamennone's Vertical Pendulum.

steel point *i* and to the lever X. The bottom of the fork has a cross-plate which is pierced by the pin *a'*. The lever X ends above in a smooth steel pin which enters the slots of the stiles, the principle being the same as that described in the Vicentini device.

The transoms S across the beams T are shaped like a

truss, as shown at QQ′, in Fig. 37, and support a small platform from which rise the columns *bb′*, united at their tops by a cross-bar. Between this bar and the platform are the axes *ss′* of the recording stiles whose sections only are seen in Fig. 37 at *ll′*.

The stiles write at *both ends*. The near ends *l l′ l″* write with ink siphons on a slow-moving ribbon of white paper, Z.

The further ends λ λ′ λ″ trace upon a quicker-moving ribbon of smoked paper, Z′. The white ribbon Z is fed by a cylinder, C, driven by a falling weight through the cord *t*, and is kept going at all times. The smoked ribbon Z′ is kept waiting until a tremor arrives, starting off a distant seismoscope, which in turn starts the ribbon.

The device for vertical motion is in principle the same as that which Dr. Agamennone uses in the small compact seismograph described in the preceding chapter, though very much more ponderous. The greatest difference between this and the Vicentini apparàtus consists in the respective devices for vertical motion. The former uses helical springs, the latter straight, flat springs.

We come now to a class of instruments of a very different character from those described, but which have in the last twenty years given some of the most important results. They are much more limited in function than the seismographs. But within their peculiar functions they are far more delicate, and may be said to take up the line of inquest where the seismograph leaves off.

The so-called horizontal pendulum is nothing more than the conical pendulum with its axis so near the vertical that the pendulum has a period of twenty to forty seconds of

time. The end of the boom, therefore, describes a circle whose plane is only a very few seconds of arc from horizontal. Its bearings are hard steel points on hard steel in order to minimise friction, and in some of them the friction of the lightest stile writing with a glass fibre-tip on smoked paper is intolerable. In place of it a very slender beam of light falling upon a mirror attached to and moving with the pendulum is reflected back upon a slowly moving ribbon of sensitised paper where it photographs its trace.

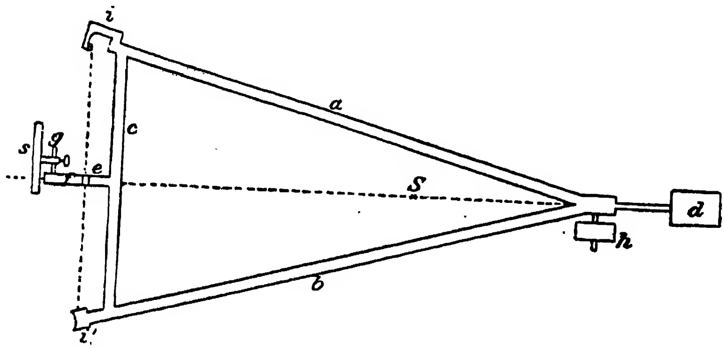


FIG. 38. Von Rebeur Paschwitz's Horizontal Pendulum.

The pendulum of Dr. E. von Rebeur Paschwitz with which some of the most valuable discoveries were made, is of this class. The pendulum itself, Fig. 38, is an isosceles triangle, of brass throughout, with the vertex prolonged into a stem. The two legs are also prolonged a little beyond the base and are provided with agate cups which bear upon hard steel points, which constitute the support of the pendulum. From the middle of the base of the triangle there projects backward an arm carrying a mirror, which receives a very slender beam of light from outside the room

and reflects it back upon a moving band of photographic paper. The mountings and adjustments, which need not be described, are of the most accurate kind, and the instrument is extremely sensitive to all causes of movement. The dis-

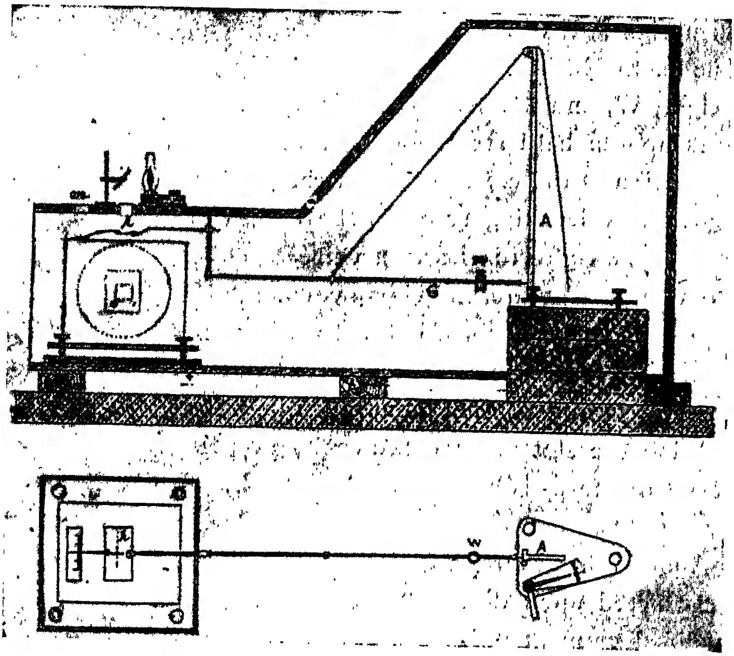


FIG. 39 Milne's Horizontal Pendulum.

tance between the points of support is 68 mm., and the horizontal length of the pendulum is 185 mm. Its weight is forty-two grammes.

Professor Milne's horizontal pendulum, shown in Fig. 39, is a very simple one. The end of the boom *p* carries a thin metallic plate with a slit in it. Underneath *p* is a box

containing a roll of sensitised paper driven by clockwork. In the cover of the box is a slit perpendicular to the one in *p*, so that a beam of light passing both slits falls as a point on the photographic paper, tracing a sinuous line as the end of the boom moves right and left. The pendulum stand and its upright, A, which is 50 cm. high, is cast in one piece. The boom which is 120 cm. (four feet) long carries a sliding weight, W, and has a quartz cup at the inner end which rests upon a hard steel point set in the upright A. The inclination of the vertical axis of the pendulum is regulated by the rear levelling screw in the base, while the two forward screws govern the transverse levelling. One of the screws has a pointer moving over a graduated arc. One degree of this arc corresponds to a known amount of displacement at the extreme outer end of the boom, and the operator knows at once how much he must turn the screw to bring the slot at that end to its proper position. The whole apparatus is housed in a wooden box to shield it from air currents.

Another horizontal pendulum is in use in Japan, and is constructed upon designs of Professor Omori. A slightly modified copy of it is now installed in the U. S. Weather Bureau at Washington, in charge of Prof. C. F. Marvin. It is shown in Fig. 40. The bob is a flat cylindrical lead weight, C, attached to a conical tubular rod, B, which terminates at B in a hard steel plug with a polished cavity to receive the point of a steel pin fixed on the main column A. The distal end of the pendulum is stayed by the wires *ww*, which unite above in a stirrup controlled by adjusting screws which regulate the height of the bob, and the inclina-

tion of the vertical axis. A band of smoked paper is applied to the drum R, and the multiplying lever L (10:1) traces upon it the lateral motions of C. The drum has a

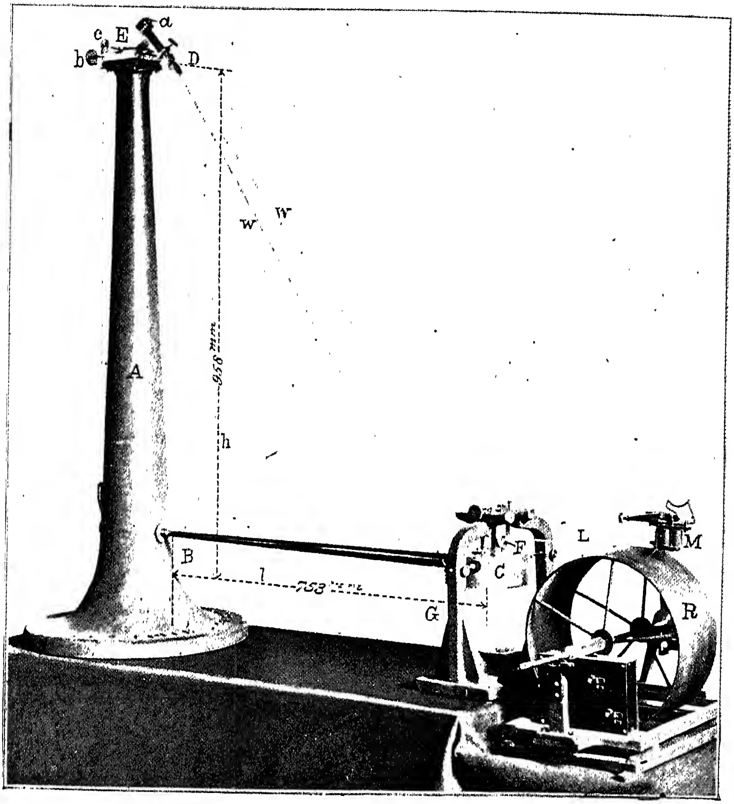


FIG. 40. Omori Horizontal Pendulum Slightly Modified.

horizontal sliding motion upon its axle, so that the lever L, when at rest, traces a helix as in any ordinary chronograph. The "ticker" M marks minutes upon the helical trace. This instrument differs from all other horizontal pendulums

in having a multiplying mechanical tracer instead of a mirror and beam of light to photograph the trace.

Decidedly the most sensitive instrument ever devised for measuring earth-tilts is the bifilar pendulum. It was suggested by Lord Kelvin and worked up by Mr. Horace Darwin. It consists, Fig. 41, of a circular mirror, M, about 20 mm. diameter, hung by a very fine silver wire made fast to two supports, PP′, one very nearly vertically over the other. Two hooks on the frame holding the mirror take hold of the loop of the wire which will hang in a vertical plane through P and P′. If the point P moves to one side, it is plain that the mirror will turn about a vertical axis; also that if P is very nearly vertically over P′, a very small sidewise movement will give a large angular movement to the mirror. The light from an illuminated porcelain disc with a vertical wire across it falls upon the mirror and is viewed through a telescope provided with cross-wires. The porcelain disc, being mounted on a slide, can be moved right or left until its wire collimates with that of the telescope and the amount of the displacement measures the rotation of the mirror.

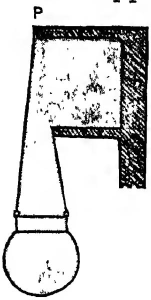


FIG. 41.
Bifilar Pendulum.

The instrument is so sensitive that it is of little use, for something is happening all the time to keep it in a state of unrest. By hanging it in glycerine the small tremors are "damped" off. But the weight of a man passing from side to side of it a considerable distance away deflects the ground, and consequently the supports, sufficiently to move the mirror. It is claimed that a deflection of the vertical

amounting to the three-hundredth part of a second of arc, *i. e.*, about one inch in a thousand miles, can be detected by this instrument. This seems incredible. In any case the instrument measures nothing but earth tilts, or deflections of the vertical, even when it is free from all other perturbations.

CHAPTER VII

SEISMIC WAVE-MOTION

Wave Concept Satisfactory but Highly Complex—Cause of the Complexity—Two Kinds of Elasticity in Solids—Elasticity of Volume—Elasticity of Shape—Defects of Homogeneity or Isotropy—Fundamental Notion of a Normal Wave and of a Transverse Wave in an Elastic Solid—The Two can be Conceived of Separately, but Must Originate Together or in Combination—But they May Separate Afterwards, Because of Different Rates of Propagation—Wave Speeds Dependent upon Elasticity-Density Ratio, which Probably Varies with Increasing Depth in the Earth—Broken or Discontinuous Nature of the Ground near the Surface Must Greatly Affect the Nature of the Vibrations—Soils Not Wholly Destitute of Elasticity—Gen. H. L. Abbot's Experiments—Professors Gray and Milne's Experiments in Japan—Surface Waves—Long-Period and Short-Period Vibrations—Gradual Recognition of Long Surface Waves at Great Distances from the Origin—Milne's Notion of Earth Pulsations—Real Character of Long-Period Waves—They Involve Horizontal Displacement without Appreciable Vertical Movement—Speculation as to the Causes of these Waves—Lord Rayleigh's Theorem of Surface Waves Not Applicable—A Fourth Class of Seismic Waves—Confined to the Epifocal or Meizoseismic Tract—Not to be Confounded with Distant Long-Period Waves—Their Destructive Character—Their Probable Origin

THE concept of a wave, or of a series of waves, in an elastic solid, be their origins what they may, is satisfactory in principle and accords so well with the actual phenomena presented in nature that we may feel confident of its verity. In its details, however, it presents many anomalies and difficulties, arising, not from doubts of the

principles involved, but from irregularities and lack of homogeneity in the media through which the waves are transmitted, and also from the higher degree of complexity in the elastic vibrations in solid media as compared with those in fluid media like air or water.

The complexity of problems in elastic solids is greatly increased by the fact that solids have two very distinct kinds of elasticity to be considered, while fluids have only one. Solids resist a change of shape, and if distorted within certain limits they resume their original shape when the distorting force is withdrawn. Fluids offer no such resistance to change of shape, and when distorted do not tend to resume their former configuration of particles. They offer resistance only to change of volume, recovering their original volume when the compressing force is withdrawn or resumes its original amount. We observe that this elastic resistance to change of volume is much greater in liquids than in gases. It exists also in solids, and in a much higher degree than in liquids. But resistance to mere change of shape without any change of bulk is peculiar to solids alone, and it is a property of a different nature from that of bulk-elasticity. The resilience of the steel spring or the torsion balance has obviously no counterpart in the properties of air or water.

Waves in elastic solids involve both kinds of elasticity, while sound-waves in fluids involve only one. The complexity of the former is, therefore, greater; for solid waves require analysis into volume-changes and shape-changes at one and the same time, and moving with different velocities. A high degree of complexity is thus met with in treating

the subject, and this complexity is manifest in the resulting phenomena themselves.

A much more serious difficulty results from the wide departure of the earth-mass from homogeneity. For purposes of wave transmission air and water may be regarded as having sensibly perfect homogeneity and perfect continuity. The case is widely different in the earth-mass. Yet we are compelled to first discuss the subject as if those properties were perfect throughout the earth, and afterwards make due allowance for the fact that they are far from being so. Most physical problems, however, are beset with many difficulties of similar nature and require, at first approach, similar assumptions.

Let us then conceive of the interior of the earth as an indefinitely extended homogeneous solid. Imagine at the depth of a few miles below the surface a spherical cavity; that within this cavity a ball of dynamite is exploded. The suddenly expanding gases may be conceived as acting with equal force at all points upon the spherical face of the cavity. The resultant direction of the forces at each point is radially outwards. Conceiving the surrounding mass as having a high degree of elasticity: every point of the surface will tend to move radially outward from the centre, imparting an outward thrust to points beyond, and propagating this thrust indefinitely. This purely ideal motion constitutes the method of movement in a "*normal*" wave. The vibration is to and from the centre of movement. It is the mode of vibration of sound-waves in air and water, as well as of *normal* waves in solids.

Now let us change the conception. Imagine some force

so applied as to tend to rotate the inner face of our cavity around a diameter as an axis. By the adhesion of the interior film to the surrounding mass this effect would be resisted in a solid, but not in a fluid. The resistance of the solid would be an elastic one. If the rotating or twisting force were withdrawn the elasticity of the material, after permitting a slight rotational displacement of the inner face, of the cavity, would at once return to its original position. But in the meantime the rotating impulse would be imparted to more and more distant envelopes, indefinitely outwards, or as far as the medium might extend.

This ideal construction may give some notion of the kind of movement in a transverse wave. It is called transverse because the vibration of the particles is across a line extending outward from the origin. It can occur only in a solid, elastic medium, and is impossible in a liquid or gas. They are sometimes called waves of distortion, as they involve change of shape, or change in the relative positions of particles, without change of volume.

In the two foregoing suppositions the effort is to present in the first case the notion of a pure normal wave, and, in the second case, of the pure transverse wave. In nature the two, though they can be conceived of separately, can hardly be separately generated, and it may be said that an elastic wave in a solid is always compounded of normal and transverse vibrations at the origin, though they may separate afterwards. It is difficult and perhaps impossible to suggest any force actually occurring in nature and acting upon the rocks below the surface in such a way as to generate a pure normal wave or a pure transverse wave. Every force which

can be suggested as at all liable to occur would create both normal and rotational displacements at the same time. A pure displacement of either kind can only be conceived of as a single or special case among an infinite number of possible ones.

That the normal and transverse waves should separate is due to the fact that their rates of propagation are different, that of the normal wave being the faster.[1] The speed of transmission of the normal wave is proportional to the square root of the ratio of volume-elasticity to density of the medium; that of the transverse wave is proportional to the square root of the ratio of the shape-elasticity (or rigidity) to the density. Volume-elasticity being greater than shape-elasticity, the speed of the normal wave is correspondingly greater.

Indicative of this separation is the fact that as the distance from the source of the tremors increases the more are they protracted, and the longer are they felt, and this, notwithstanding the energy of the vibration grows rapidly less with distance. The tremors seem to be longer drawn out, as if a portion of them lagged behind and got more and more behindhand the farther they travelled from their origin. An earthquake which is over in a minute or less, near its centre, where its action is most energetic and destructive, may last five or six minutes a few hundred miles away where its vibrations are much feebler; and at still greater distances, where delicate instruments are required to disclose its movements,

[1] This separation of the two kinds of waves is only an inference from the general theory of wave-motion in elastic solids. It has never been directly observed, though long-distance observations strongly support the presumption.

it may last a large portion of an hour, or even several hours. This would hardly be a common or characteristic phenomenon in all great earthquakes if all the tremors travelled with the same speed.

As deeply as we can observe them the rocks are very heterogeneous in character: *i. e.*, for the purposes of our discussion they vary in density and elasticity; and below the reach of our observation the few indications we have suggest that these variations continue into the indefinite depths of the interior. Near the surface the rocks are very much "jointed," cracked, and fissured, and the strata consist of materials which are not only heterogeneous among themselves, but their elasticities are different in directions across the bedding from what they are parallel to the bedding. As the depth increases the increasing weight and pressure of the superincumbent mass tends to keep all joints and fissures closed up until, at a depth of a very few miles, open spaces or discontinuities become impossible. But differences of density and elasticity due to differences in the constituent material may persist to indefinite depths below.

It is rather discouraging to ask what inferences we are justified in drawing from volcanic phenomena respecting the condition of the earth's interior. The extent of liquid fusion is a question upon which we have little to guide us. The old view that beneath a comparatively thin crust the great interior regions of the earth are everywhere in a state of fusion has latterly been antagonised by the view that the interior is, in the main, solid from centre to circumference, and that the lavas are erupted from comparatively small isolated vesicles near the surface. And this later view is

much more in accord with the general condition of facts than the older one. Just here, however, we are concerned with it only in so far as it is related to the properties of the earth-mass as a medium for the transmission of elastic vibrations.

According to the undulatory theory, wherever a vibration reaches a bounding surface it undergoes reflection. When it reaches a medium of different density it is refracted. And since waves in the earth cannot travel far without meeting one or the other, they are quickly broken up into secondary and tertiary waves and derivatives of more and more remote order. Some analogy may be suggested by the calm surface of a pool of water when a pebble is cast into it. The symmetrical waves expand until they reach the border. They are reflected back until the surface is covered with a tremulous shagreen of wavelets in which all symmetry is lost.

Close to the surface the ground is so far from being a homogeneous, elastic solid, that it becomes necessary to modify profoundly the fundamental conception of an elastic wave. The soils, the alluvia, the sands, which rest upon the surface, the shattered, shaly strata, broken by frost and weathering into a mass of disjointed rubble, receive the impulses from the more coherent masses below, but the characters of the vibrations are greatly changed. The energy of the elastic waves from beneath is expended in imparting to these incoherent surface layers a new set of motions differing much from the original vibrations. These new motions may still be oscillatory, but the material having, as a mass, very little elasticity, it cannot take up the motion

and propagate it in its original wave lengths or amplitudes.

It would be an error, however, to treat soils and alluvia as wholly destitute of elasticity and of the capacity to transmit elastic waves. Though they are not a continuous mass of material they are composed of small fragments in contact. We have only to consider the various particles or grains as being in contact each to each at one or more points to perceive that one may convey to another a thrust or impulse, and thus have a capacity to transmit it, though with a much lower degree of elasticity than in the case of a continuous solid mass. The density, too, would be less because of the interstices, but not in so great a ratio. Therefore the speed of propagation would be less. This has been shown by many experiments with falling weights dropped upon the soil or sand, and the shock of the impact noted at a distance, together with the time of transit. The speed of transmission is always found to be, in sands and soils, a few hundreds of feet per second, while in the depths of the earth the speed is as many thousands of feet per second.

The seismic action which is visible at the surface, and whose results can be studied, usually consists of movements of the less coherent materials of the soil and alluvia, while the true impulses which generate these movements are perceived only by the imagination guided by mathematical analysis. Ordinarily the true seismic vibration is transformed in passing into the surface layers. The resulting movements of the soil are largely influenced by accidents of the ground, and these vary so much in kind and degree that it is hardly practicable to embrace any considerable part of

them in any one formula or mode of treatment. The capacity of soils and sands to transmit impulses in the manner of elastic waves had long been believed when Mr. Robert Mallet in 1851 laid before the British Association for the Advancement of Science a series of experimental results, wrought out in a most careful manner, demonstrating the reality of that view. In 1874, Gen. H. L. Abbot, of the U. S. Engineers, took advantage of an opportunity afforded in the blasting operations at Willet's Point by the explosion of 50,000 lbs. of dynamite to investigate the propagation of the earth tremors caused by it.

One of the results of Abbot's experiments indicated a much higher speed of propagation than had before been conceived as possible. Mallet's results had given:

Velocity in sand	825 feet	per	second
Velocity in much-shattered granite	1306 "	"	"
Velocity in more solid granite	1665 "	"	"
Velocity in quarries at Holyhead	1320 "	"	"

Abbot's first results at Willet's Point indicated about 8300 feet per second. Mr. Mallet appears to have suspected some important errors of observation in this experiment and so expressed himself. Gen. Abbot, therefore, determined to avail himself of explosions of dynamite which were made from time to time at the torpedo station at Willet's Point for experimental purposes, and these subsequent experiments not only confirmed the first results, but developed some facts which had not before been brought to light. He had been led to query whether the speed of transmission might not increase with the intensity or power of the original shock, and also whether the time of arrival might not be

made perceptible earlier by the use of more delicate instruments. A basin of mercury viewed through an inverted telescope was used, and the tremors imparted to the mercury furnished the signal of their arrival. By increasing the magnifying power of the telescope from six to twelve diameters the observed time of the first appearance of the tremors was very materially hastened. The time during which the agitation of the mercurial surface was recognisable was also notably increased. The following table shows these interesting results:

No. of Obs.	Date	Cause	Distance (Miles)	Telescope [1]	Time of Transit	Tremors	
						Duration	Speed: ft. per sec.
1	Aug. 18, 1876	200 lbs. dynamite	5.±	B	5.±		5280±
2	Sept. 24, 1876	Hallet's Pt. Exp.	5.184	A	7±	63.±	3873±
3	" " "	" " "	8.330	B	5.3	72.3	8300
4	" " "	" " "	9.33	A	10.9	23.5	4521
5	" " "	" " "	12.77	B	12.7	19.	5309
6	Oct. 10, 1876	70 lbs. powder	1.36	A	5.8	inst.	1240
7	Sept. 6, 1877	400 lbs. dynamite	1 17	A	1.8	7.8	3428
8	" " "	" " "	1.17	B	0.7	17.8	8814
9	" 12 "	200 " "	1.34	A	1.05	8.8	6730
10	" " "	" " "	1.34	B	.81	17.1	8730
11	" " "	70 lbs. powder	1.34	A	1.27	4.8	5559
12	" " "	" " "	1 34	B	.84	15.1	8415

In 1880 Messrs. Gray and Milne conducted a series of experiments in which the tremors produced in the soils by falling weights were observed. The drop weighed about a ton and was allowed to fall from heights varying from ten to

[1] A indicates that the arrival of the tremors was noted by a telescope magnifying six diameters; B, a telescope of twelve diameters.

thirty-five feet. At varying distances the vibrations were received on a seismographic plate where they were recorded automatically. The record seemed to indicate that both normal and transverse vibrations occurred. One set of motions appeared to be normal in mode, *i. e.*, to and from the origin of the impulse. These were very quickly followed by another set apparently transverse, *i. e.*, across the line from the origin.

Thus far we have spoken only of certain ideal waves in an elastic solid. They are of two kinds and the two simplest kinds, viz.: the normal and the transverse. They are contemplated as expanding radially outwards from a central point or origin and propagating themselves through a medium which has no other limit than the surface of the earth. Recent progress of seismologic investigation has brought to light a third class of vibrations called surface waves, which have awakened great interest among students of earth physics.

It may be questioned whether they are in reality a new discovery or anything else than a more definite and distinct recognition of a form of seismic motion which has always been imperfectly recognised as far back as the time of Thales and Aristotle, and only very recently brought into lines with the most modern concepts of wave-motion. All writers ancient or modern seem to have been aware that seismic motion is vibratory and that there are vibrations of long periods and others of short periods. The ancient philosophers, Thales, Aristotle, Seneca, Pliny, Pausanias, seem to imply the idea that during an earthquake the ground moves in waves like the sea, while subject at the

same time to sharp blows or shocks from beneath causing a "successatory" or quick up-and-down motion superposed upon a slower undulatory movement. Thus Seneca says: "Duo genera sunt, ut Posidonio placet, quibus movetur terra. Utriusque nomen est proprium, altera succussio est, cum terra quatitur et sursum ac deorsum movetur, altera inclinatio, qua in latera, nutat navigii more."[1] The notion that in an earthquake the ground undulates like the surface of the sea is surely very ancient and was deeply planted in the minds of the inhabitants of countries most subject to quakes, in Greece, the Ægean Archipelago, and southern Italy. It also became the foundation of the idea that the earth was composed of a crust floating upon a liquid.

The more detailed accounts of earthquakes in the literature of the nineteenth century had often recited the occurrence of a swinging motion like that which is felt on the deck of a ship, though less in amplitude. It was an invisible motion, but it was distinctly felt and sometimes caused nausea. At considerable distances from the epicentrum people in the upper stories of buildings felt a swinging motion, and the chandeliers or other freely suspended objects would swing, while no movement would be felt by those out of doors or upon the ground floor. This slow, swinging motion would be perceived when no other form of vibration was sensible. It was observable not only at great distances from the epicentrum but at intermediate points.

When the seismograph came into use in Japan the first

[1] Seneca, *Naturales Quæstiones*, lib. vi., cap. 21.

records obtained from it showed that vibrations of short period, from a tenth to a quarter of a second, were usually superposed upon and simultaneous with vibrations of one or two seconds. The short vibrations were usually a little before the longer ones, but quickly died away, leaving the longer ones predominant and at length the exclusive form of oscillation. Thus was given a partial confirmation of the ancient idea as expressed by Seneca. The seismograph showed the long, swinging movement like the rocking of a ship and at the same time the quick tremors which, though not strictly a *succussio*, were enough like it in their effects to be readily mistaken for it.

The continued use of the seismograph in the observatories of the seismologists who were so earnestly and ably studying their records in Japan soon brought to light vibrations of still longer periods, whose significance was not at first fully understood. Periods of three or four seconds were sometimes observed. But they were from earthquakes originating in distant parts of the islands or at sea, and presumably the vanishing vibrations of distant quakes of considerable power. Occasionally records would be obtained of vibrations having periods of one, two, or even three minutes, and having no apparent association with any earthquake whatever. Similar observations were made in Italy with the great vertical pendulum seismographs soon afterward introduced. For a considerable time no one was bold enough to attribute these long-period vibrations to distant earthquakes. There were two objections which seemed to negative that explanation: (1) The earthquakes which might produce vibrations forcible enough to affect even a

delicate pendulum three or four thousand miles away must necessarily be of great power and could not escape notice and world-wide celebrity unless occurring in localities very far from human observation or perhaps in the depths of mid-ocean. But these long-period oscillations could seldom be identified with any reported earthquake. (2) The periods, sometimes exceeding two minutes and in most case ranging from twenty to sixty seconds, seemed to imply a class of disturbances of very different nature to those recorded by seismographs in near-by earthquakes whose longest periods seldom exceeded three seconds. To accept them as the distant manifestations of these waves of two or three seconds' period would involve a remarkable lengthening of the period, with increasing distance from the origin, for which no warrant could be found in the known laws of wave-motion.

The leading seismologists, therefore, with commendable caution refrained from attributing them to distant earthquakes until more evidence could be collected. In the meantime Prof. Milne gave them the non-committal name of Earth Pulsations and indulged in some ingenious and interesting speculations as to their true nature and origin.

The steadily increasing number of observing stations equipped with sensitive instruments in many parts of the world at length accumulated sufficient evidence to establish the principal facts relating to these long-period waves, though some important questions are still in the problematical state. As these waves are to form the subject of a future chapter it must suffice for the present summary view to state the principal facts with the greatest brevity.

(1) These waves originate at the centra of powerful

earthquakes and are propagated upon the earth's surface to great distances, even to the antipodes and beyond. To what depths the earth is affected by them is wholly unknown. It is presumably some considerable fraction of the wave-length, which often exceeds fifty kilometres and is known in some cases to exceed a hundred kilometres. (2) The wave-lengths and periods appear to increase as the disturbance travels. At the epicentrum their periods may be in most cases from one and a half to three seconds and their wave-lengths from four to eight kilometres. At a distance of six thousand miles, or one-fourth the earth's circumference, the periods may be from twenty to forty seconds and the wave-lengths may be as great as a hundred kilometres. (3) The speed of propagation, while it probably is not strictly constant, does not appear to vary much. The records are doubtless more or less affected by the varying degrees of sensitiveness and promptitude with which different instruments respond to these disturbances. But while full allowance for instrumental differences might account for much of the inequality of speed in the waves, it cannot account for all of it, and there is at present good evidence that the speed really varies within very moderate limits, both with respect to the waves of different quakes and with respect to different parts of the paths of waves from the same quake. Their usual rate of propagation is from two and a half to three kilometres per second. (4) There is a remarkable uniformity in the periods of these surface waves from any one quake, and at any one place of observation, though the periods of different quakes may be different at the same instrument, and the periods of the same quake may be different at dif-

ferent distances. This cannot be due to the instrument itself, which has its own period and one which seldom coincides with that of the earth-vibration.

A question has lately arisen as to what is the real character of the motion in these surface waves. At first it was taken for granted that the waves were, in effect, so far similar to very long, flat waves in a body of water that they had crests and troughs and caused alternate rising and falling of points in the surface of the ground, though the height from trough to crest (double amplitude) might be only a few centimetres, while the length from crest to crest might be many kilometres—even a hundred kilometres or more. Nor would this great disproportion between amplitude and length be unexampled or at all extraordinary. The tidal wave is as long as the ocean is wide and only two or three feet high. The great sea-waves generated by earthquakes are often much more than a hundred miles long and only three or four inches high. If this concept of the waves were correct, then the surface must be subjected to a tilting motion as the waves pass, now to one direction and then to the opposite. The tilt must be a very minute one, but easily within the capacity of a horizontal pendulum to show it. Thus a tilt of one foot in forty miles would be equivalent to a tilt of one second of arc. All standard pendulums for seismologic investigation are expected to be sufficiently sensitive to respond to a tilt and consequent deflection of their axis of oscillation of considerably less than a second of arc. It is claimed for the bifilar pendulum that it will show a deflection or tilt of one three-hundredth of a second.

After some years of observation, and after hundreds of

these long-distance disturbances had been recorded at the various stations, a question arose whether the displacement of the pendulum was due to a tilting as at first believed, or was caused by a horizontal movement of the ground, leaving the pendulum to act as a steady-point by its own inertia while the ground moved to and fro. It was noticed that instruments which by their structure ought to be highly sensitive to tilts and only slightly sensitive to horizontal displacements often failed to indicate a disturbance where instruments less sensitive to tilts were strongly affected. In general, the action of most instruments could be explained in most cases by either form of movement, though there were some instances where the records were more indicative of horizontal movement of the ground than of tilting.

To test the matter more decisively Dr. Wilhelm Schlüter devised an instrument which he termed a Clinograph. Its distinctive feature was its high sensitiveness to tilts and insensitiveness to horizontal displacement. The result of a long series of trials was that his instrument gave no indication whatever of surface waves while two Rebeur-Paschwitz pendulums installed in close proximity to it gave twenty seismograms of unmistakable character.[1]

Dr. Schlüter, however, did not conclude that there was absolutely no vertical component in the motion and therefore no tilting whatever, but merely that if there had been any such tilting it was too small to affect the photographic trace of the clinograph. He was confident that the only motion present in appreciable amount was horizontal.

[1] "Schwingunsart und Weg der Erdbebenwellen," von Wilhelm Schlüter. *Beiträge zur Geophysik*. Bd. v.

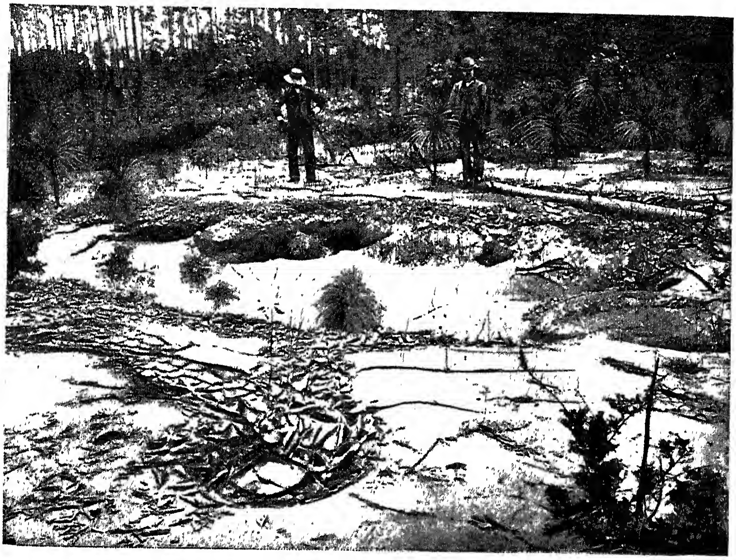


FIG. 42. Craterlet from which Large Quantities of Water and Sand were Ejected in Charleston Quake.

FIG. 43. Railroad Distorted by the Charleston Quake.

Prof. Milne who in common with the great majority of leading seismologists had at first believed that these pendulum seismograms were produced by tilting, at length reached an opinion very similar to Dr. Schlüter's. In the British Association Report for 1902 he sums up the considerations for and against the original view that the movements of the horizontal pendulums are due to tilting. He says:

OBSERVATIONS ADVERSE TO TILTING

"1. Clinometers have hitherto failed to detect any tilting effects.

"2. If it is assumed that the records of horizontal pendulums give angular values for tilting, and from periods of waves causing those tiltings, and the velocity with which those waves are propagated, on the assumption of simple harmonic motion, we calculate their length, we have all the elements which are required to calculate the heights of these waves. Now these heights (as calculated) are frequently as much as one or two feet and apparently represent accelerations $\frac{1}{50}$ of gravity. The magnitude of these quantities is certainly sufficient to create a suspicion that the angular values assigned to large waves has hitherto been exaggerated.

"3. The slight evidence of vertical displacements afforded by experiments for this purpose in Japan and in the Isle of Wight.

"4. Dr. Omori's observation that the amplitude of seismograms is not dependent upon the sensibility of the seismographs to tilting, suggests that the movements represented by large waves are horizontal rather than undulatory.

"5. The smallness and paucity of records obtained from bifilar pendulums."

OBSERVATIONS SUPPORTING THE UNDULATORY OR TILTING HYPOTHESIS

"1. Surface undulations exist in epifocal districts, and these, by the movement of water in ponds and lakes, the movements of

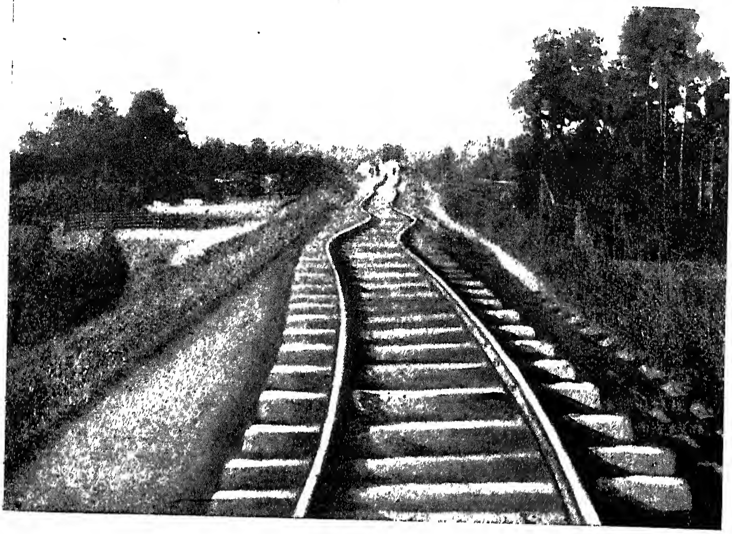


FIG. 44. Railroad Distorted by the Mino-Owari Quake in Japan.

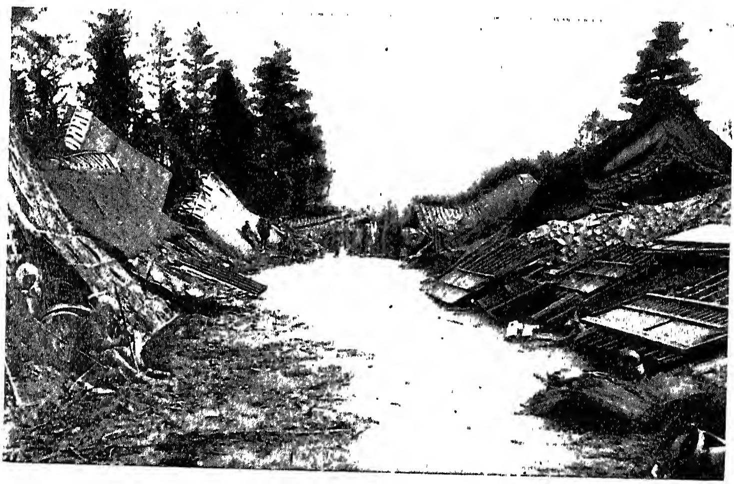


FIG. 45. Village Wrecked by Japanese Quake.

bubbles in spirit levels, the apparent movements of stars in the field of telescopes, and by other phenomena, have been detected in districts many hundreds of miles beyond the epifocal area.

"2. The approximately constant velocity of propagation assigned to large waves.

"3. Observations which show that the apparent magnitude of a seismogram is dependent upon its sensibility to tilting. This conclusion is apparently contrary to that arrived at by Dr. Omori.

"4. The indications of a vertical component of motion which have been recorded."

It is not apparent that there is a necessary conflict in the two classes of considerations. Not one of them is decisive in itself and none can furnish more than a probability which is greater in some than in others. Collectively, however, the various groups of observations argue strongly for the view which Dr. Schlüter and Prof. Milne both adopt, viz.: that the chief motion is horizontal, and at long distances, say 90° or more, the vertical component, though it may exist, is too small to be picked up by the clinometer though it may sometimes be detected by the bifilar pendulum. In the epifocal district the long waves are surely undulatory and the vertical component can be measured. But even there, where it has its maximum value, the vertical component is but a small fraction of the horizontal. As the wave recedes from the epicentre the vertical component rapidly diminishes. But the horizontal component, being distributed over a steadily increasing wave-length, diminishes at a slower rate. At a great distance from the epicentrum the vertical component is lost or insensible while the horizontal component still persists.

The causes, or motor forces, of these waves have been

FIG. 46. Fissures Opened, and Wrecked Railway Bridge, Mino-Owari Quake.

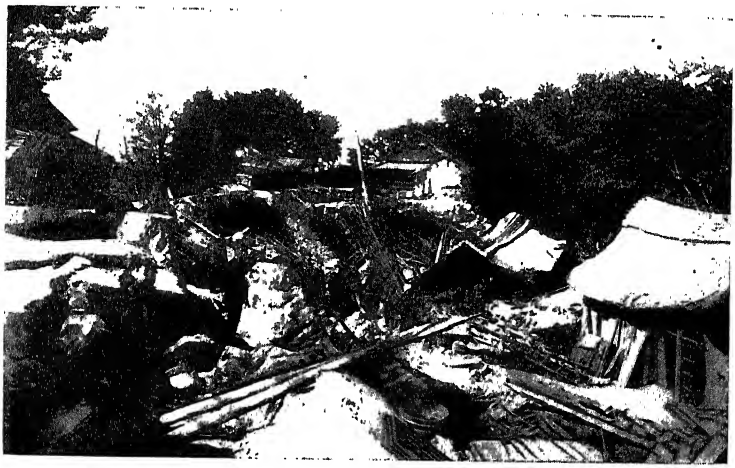


FIG. 47. Village almost Engulfed, Mino-Owari. The Land is Reported here to have Slipped about 20 Metres.

the subject of much speculation. For a time much attention was given to a theorem by Lord Rayleigh,[1] published in 1885, on the action of waves propagated along the plane, free surface of an infinite, isotropic, elastic solid. Lord Rayleigh discusses two cases, one in which the medium is incompressible, *i. e.*, having only elasticity of shape as is assumed to be the case in the ether, the other in which the medium is endowed with both kinds of elasticity. The results are similar in both cases. The disturbance in a direction parallel to the free plane surface penetrates only to a depth of about one-eighth of a wave-length in the incompressible medium and of about one-fifth of a wave-length in the other case. The motion of a particle at the surface is in an elliptic path, the major axis of the ellipse being vertical.

This motion of the surface particles would thus give rise to a wave similar to a wave in water, though the potential energy involved is elasticity instead of gravity. But it seems impossible to use Lord Rayleigh's theorem in the present problem, as the elliptical motion with the major axis of the ellipse normal to the surface gives us a vertical component whose effect upon the form of the wave surface is equivalent to a tilting up and down alternately, according to the phase of the wave, and this is so strongly opposed by observation that we seem to have no alternative but to dismiss it.

There is still a fourth class of earthquake waves, and to the inhabitants of badly shaken regions they are doubtless of the most importance. The other kinds of waves, the

[1] *Proceedings London Mathematical Society*, vol. xvii.

FIG. 48. Wooden House Wrecked by Charleston Quake, near Summerville, S. C.

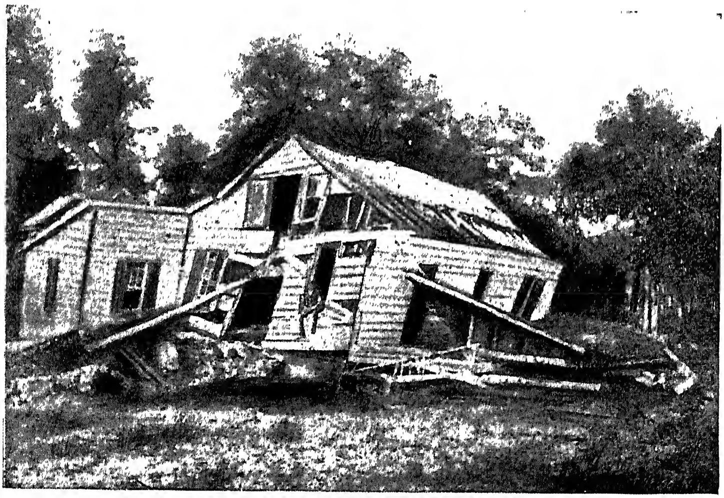


FIG. 49. Wooden House Wrecked by Charleston Quake, near Summerville, S. C.

normal, the transverse, and the surface waves, are seldom seen though they are more or less felt. The fourth kind are both seen and felt, and are often the most conspicuous and terrifying. They are also the most destructive. They occur only in the epifocal districts of the great earthquakes, and never far outside of them.

In most of the greater or very destructive quakes the ground is seen to be traversed with swiftly moving waves. Their outward forms as described by eye-witnesses exactly resemble flat waves on the water. They move with a speed which to the observer seems no doubt very swift, though in reality it cannot be more than a small fraction of the speed of the waves already mentioned. But they are seen to tilt buildings, to raise and lower the pavements of streets, to swing and lash trees and telegraph poles, to hurl down in ruins brick or stone walls, to sway or even snap off tall chimneys. The ground has been seen to open in cracks in the crests of the waves and close together in the troughs, squeezing out water thick with sand and mud, which is spurted upwards or which forms little craters around the vent holes. The ground if soft and unconsolidated is often permanently distorted by these waves. Railroad tracks and road-beds partake of this distortion, which is often extreme. Excessive distortion was suffered by the road-bed and rails of the South Carolina railway where it crossed the epicentral trail of the Charleston quake. Similarly the Japanese railway in the Mino-Owari quake, and the Assam railway in the great convulsion of 1897, were so distorted that entire reconstruction was necessary.

It is these waves which produce cracks in the soil and

unconsolidated alluvia, which remain after the shaking is past. They also cause landslips by forcibly shaking the materials with aid of gravity down the slopes of the bed rock. It is a frequent occurrence for the alluvial banks of rivers to crowd during the vibrations towards the middle of the stream, opening long, wide cracks on the banks. The

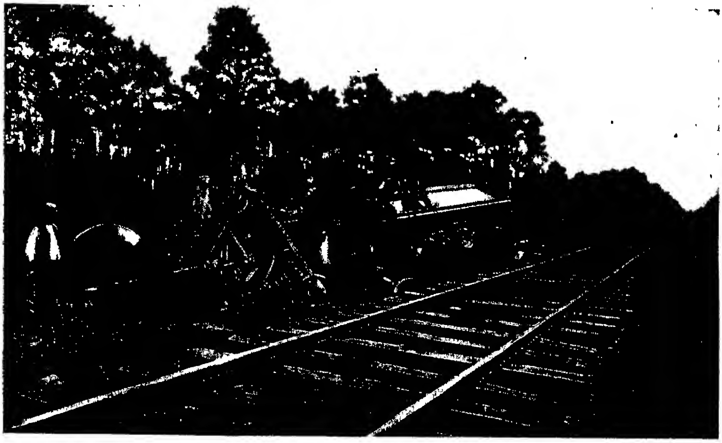


FIG. 50. Locomotive Derailed while Running near Epicentre of Charleston Quake. This and the Preceding Disasters from Figs. 42 to 49 are All the Results of Epifocal Waves.

sliding alluvia carry the abutments of bridges with them, crushing up the spans or trusses as if they were wax.

For a long time these motions of the ground were questioned by seismologists, who could hardly believe that such waves could be visible. But the testimony of hosts of credible witnesses at length reached such force and volume, was so concordant and circumstantial, and was so universal in all great earthquakes that it overwhelmed all doubt or

dispute about the fact and left as the only questions the nature and causes of these waves.

The three preceding classes of waves (normal, transverse, and surface waves) all have as their potential energy of motion the elasticity of the materials composing the earth-mass. There is abundant reason for inferring that these epifocal waves are of a very different nature and have no relation to elasticity. Their lengths are too small, their amplitudes too great, and their speeds of propagation too slow to be dependent upon elasticity. Their lengths indeed are not known exactly, but the inference from the accounts given of them is that their wave-lengths range between ten and fifty metres and their heights between five and thirty centimetres. Their speeds of propagation are even more uncertain, but are almost surely less than one hundred metres per second. These quantities do not fall within the possible range of elastic wave-motion in any such material as the rocks or soil.

That they are the results of deeper waves passing from a highly elastic to a feebly elastic medium is obvious, and the result must be to diminish the speed of propagation and also the wave-lengths through the yielding and deformation of materials thus invaded. That a greatly increased amplitude should result from the much lower elasticity might also be expected. Probably the deeper transverse waves are the principal agents of this motion. They may be conceived of as flinging the loose, discontinuous surface layers right and left with resistless energy, finding here and there a spot which can be made to oscillate with a period which is some small multiple of their own.

CHAPTER VIII

AMPLITUDES AND PERIODS

Meaning of Amplitude—Illustrations of it Especially Near an Epicentre—Amplitude of Vertical Motion—Professor Sekiya's Investigations of Vertical Amplitudes and Periods—Lengthening Out of Periods as Distance Increases—Illustrated in Charleston Quake—Increase of Wave-Length with Distance—Effects of Variable Ground upon Amplitudes and Periods—Milne's Results in Observing them at Great Distances from their Origins — Omori's Investigation — His Subdivision of the Phases—Long Distances and their Effects on these Elements—Omori's Conclusion Respecting the Effect of the Free Period of the Ground—Duration of Quakes and its Variation with Distance

BY amplitude is meant the distance which a vibrating particle moves from its mean position of rest. It should not be confounded with double amplitude, or the sum of the distances on both sides of its mean position to which it may move. Thus the amplitude of a water wave is the height either above or below the mean water-level to which a particle rises or falls, and the height from crest to trough is the double amplitude. It is impracticable, or at least inconsequential, to consider amplitude apart from period, which is the time required to complete an oscillation. For wave-motion is a matter of energy in which time is an inseparable element. The energy of the quake is proportional to the square of the velocity of the earth particle and this velocity is in turn proportional to the amplitude directly

and the time of its excursion, or period, inversely. With equal amplitudes the effects are inversely proportional to the square of the period. Or with equal periods they are proportional to the squares of the amplitudes.

In an ordinary earthquake we have, near the epicentrum, a wide range of both amplitudes and periods, from the smallest quivers to the longest swings. The smallest period which any seismograph can register is not far from the twentieth of a second, though in most instruments the records seldom show a smaller interval than a tenth of a second. This is not because there are no quicker tremors, but simply because the structure of most instruments does not admit of any legible response to quicker ones.

As regards the greatest amplitudes of earthquakes they are never recorded. They are sufficient to destroy or overthrow the instrument and thus prevent its record. It is certain, however, that in some cases it is as great as a foot and possibly in extreme cases a little more. But such an amplitude occurs only on soft ground near the epicentre of a quake of the highest order of energy. Even in such a quake the amplitude on firm rock is probably never more than two inches. An amplitude of .5 mm. and a period of .2 second is quite sensible to the feeling. An amplitude of 10 mm. and .5 second period is a decidedly forcible quake, sufficient to crack walls badly, while 20 mm. and .5 second is a destructive one.

Generally the amplitudes and periods both increase together near an epicentrum, but the former increases more than the latter, thus giving a greater acceleration for the larger swings and consequently greater destructiveness.

Amplitudes of 50 and 60 mm. are not exceptional in great quakes, but their periods are usually from 1.5 to 2.5 seconds.

Thus far amplitudes of horizontal motion have been alone considered. Ordinarily in an epicentral tract vertical motion is also observed, though sometimes it is so small relatively to the horizontal that it escapes detection. Vertical vibration was made the subject of careful investigation by the late Prof. Sekiya of Tokio, who analysed the seismograms obtained during the years September, 1885, to September, 1887.[1] In this study Sekiya ascertained the following quantities: (1) the greatest displacement ($2\ r$) (amplitude) of the ground in each shock; (2) the complete period (t) of each; (3) the maximum velocity (v) of the earth particle or $v = 2\, r\, \pi\ /\ t$; (4) the maximum acceleration $\frac{v^2}{r}$; (5) the direction of the maximum horizontal motion of the ground; (6) the duration of the quake; (7) the distance and direction from Tokio of the origin, and also the area of the disturbed region.

In most of the observed quakes no vertical motion appeared, that is, the ground moved only in a horizontal plane. This, however, was because the quakes which disclosed no vertical motion originated at a great distance from the seismograph. Rejecting doubtful cases and those noted as faint tremors, only twenty-eight out of one hundred gave clear indications of vertical movement. Taking account only of those in which it was observed it appeared that the vertical amplitude averaged about one-sixth of the horizontal, and the vertical period about one-half of the horizontal, or, more precisely, as 1 to 1.8. The vertical motion

[1] *Trans. Seismological Society of Japan*, vol. xii., p. 83, 1888.

began in the earliest stages of the quake but ended before the horizontal movements, the average ratio of the respective durations being as 1 to 3. The vertical motion almost invariably appeared when the horizontal motion had reached 1 mm., which was more than the average amplitude in ordinary quakes. Out of one hundred shocks there were eighteen cases in which the ground moved more than 1 mm. Out of these eighteen cases vertical motion occurred in fourteen and did not appear in the other four. On the other hand, vertical motion appeared in some cases where the horizontal motion was less than 1 mm.

In earthquakes showing both horizontal and vertical motions feeble but quick period tremors of both types simultaneously preceded the principal movements. The more decided and pronounced motion usually appeared first in the horizontal component, and then came the large vertical movements.

From the general fact that the vertical amplitude is invariably less than the horizontal, while the vertical period is also less but not in so large a proportion, it follows that the vertical acceleration is always less than the horizontal. It does not appear that any horizontal acceleration has ever been recorded so great as that of terrestrial gravitation. Consequently much less has any been recorded sufficient to actually project upwards any loose object resting upon the ground. Whether an acceleration as great as gravity could be produced under any circumstances is a question too indefinite in its terms to discuss. But it can at least be affirmed that no earthquake has ever been proven to have exerted so great a force, nor is it even in a very small degree

probable that any has ever exerted a vertical force half so great—unless indeed we are to attach full faith and credit to Humboldt's account of the Riobamba earthquake.

Although this famous incident has been made to do duty in most of the text-books of the last century it is regarded here as largely mythical. It does not appear that Humboldt scrutinised with sufficient care the grossly exaggerated accounts of it given him. If he had put those stories to the test of a little mechanical philosophy, he would have seen that an acceleration sufficient to project human beings across a wide ravine, or a hundred feet in air, would have projected soil, rocks, and everything else along with them.

In the epifocal tract of an earthquake, vibrations of all periods, amplitudes, and wave-lengths seem to come at once and at the very beginning of the disturbance. Or, if there is any difference in the time of the respective arrivals, it is very small. At Summerville the Charleston quake of August 30, 1886, had been preceded by occasional heavy booming sounds like great cannon fired in the distance or like far-off quarry explosions. But as they passed without any result they became at last familiar and lost in some degree their alarming character. When the final crash came it was unheralded. As one of the witnesses described it: "I had been out in the garden admiring the beauty of the evening and was entering the door of the hall of my house, when, without any rumble or warning, the floor seemed to sink under me. I seized the door-jambs to steady myself, when the floor seemed to go down in front of me at an angle of twenty-five or thirty degrees. It was so sudden and unexpected that I was thrown forward into the hall about ten

feet and as quickly thrown backwards, and before I could fall upon the piazza I was again thrown forward into the house." The accounts from the near vicinity of the epicentre of that earthquake all indicate that no minor vibrations ushered in the greater movements.

The descriptions of witnesses in Charleston, which is about twenty or twenty-three miles from the epicentre, show a notable difference, as the following account of the editor of the principal newspaper will indicate:

"At the time of the first shock the writer's attention was vaguely attracted by a sound that seemed to come from the office below, and was supposed for a moment to be caused by the rapid rolling of a heavy body, as an iron safe or a heavily laden truck, over the floor. Accompanying the sound there was a perceptible tremor of the building, not more marked, however, than would be caused by the passage of a car or dray along the street. For perhaps two or three seconds the occurrence excited no surprise or comment. Then by swift degrees, or all at once—it is difficult to say which—the sound deepened in volume, the tremor became more decided; the ear caught the rattle of window-sashes, gas-fixtures and other movable objects. . . . The long roll deepened and spread into an awful roar, that seemed to pervade at once the troubled earth, and the still air above and around. The tremor was now a rude rapid quiver that agitated the whole lofty strong-walled building as though it were being shaken—shaken by the hand of an immeasurable power, with intent to tear its joints asunder and scatter its stones and bricks abroad, as a tree casts its over-ripened fruit before the breath of the gale."

This account, which is representative of many concordant descriptions in the city of Charleston, plainly differs from those given at Summerville. It indicates that the vibrations which threw down walls and wrecked streets, and which

must have had relatively large amplitudes and long periods were preceded by rapid quivers of small amplitude and period. The time interval between the arrivals of these two classes of vibrations was short indeed—perhaps four, six, or at most eight seconds. But it was perceived by many observers even in the midst of a destroying earthquake and it was unmistakable. Near the epicentre no such separation or distinction of vibrations was noted. All came together, big and little, long and short, rapid quivers and slow swings, though in the terror and confusion only the greater movements impressed themselves on the memory.

Proceeding a stage farther away from the epicentrum, say to Columbia, S. C., or Savannah, each about seventy-five miles away, we find these differences in the times of arrival of short and long vibrations still greater. At those distances the short tremors seem to have nearly died out. Few noticed them, but those few observed that a very considerable interval passed before the arrival of the longer pulsations which cracked walls and threw down chimneys and plastering.

Proceeding still farther away, say three hundred to four hundred miles from the epicentrum, the short vibrations were no longer perceptible to anybody nor under any conditions, though a good seismograph might still have caught them if there had been any such instrument on duty anywhere east of the Mississippi. But there was none. All that was felt was a long, slow swing—much slower than the longest swing at Summerville or Charleston; perhaps five or six times as long (in duration and not in amplitude).

The facts here suggested are general. As the distance

from the centre of dispersion increases, the greater become the periods and wave-lengths of all kinds and the longer is the duration of the entire period of disturbance. The amplitudes alone decrease and these at a less rapid rate than might have been expected. The various forms of vibration seem to break up into waves which have different speeds of propagation. At distant localities the swifter forms arrive first and the interval between their arrival and that of the next class increases with the distance. Each wave grows longer in period as it speeds along. The short, sharp quivers in the epicentral tract having five, ten, fifteen pulses in a second have periods of two or three seconds five hundred kilometres away, if they have not lost on the road so much of their energy that sensitive instruments cannot detect them.

This general fact is indicative of the transmission of the vibrations through a medium that is not isotropic, or whose elasticity is imperfect. Suppose in the propagation of a transverse wave in a solid the wave front reaches a position where the elasticity is slightly less than that in previous parts of its course. The particles, moved right or left, will be urged a little farther than before because of diminished elastic resistance, *i. e.*, their amplitudes will be slightly increased. But since the elastic force is less they will be slower in returning to their original places; *i. e.*, their periods will be increased also. The same reasoning would hold good for normal vibrations.

This change of periods and amplitudes is well illustrated in the epicentral tracts of all moderate quakes where the ground is heterogeneous, consisting partly of rock, partly

of soils of varying firmness. The amplitude is invariably much greater in the soft ground than in the hard, and the period is longer. There has been a passage of the waves from the harder and more elastic rocks below into the softer and much less elastic soils above. The longer period would offset the greater amplitude if the elasticity were perfect and thus leave the amount of energy unchanged. But the elasticity being very imperfect, some of the energy is lost or dissipated in producing deformations of the ground so that the amount of energy diminishes.

If the elasticity of the earth as a medium of wave propagation were perfect the theory indicates that the amplitude should decrease in a simple inverse ratio to the distance, while the period should remain constant. This would be equally true of both normal and transverse waves. As a matter of observed fact the period always increases with the distance, while the amplitude diminishes less rapidly than in an inverse simple ratio. The difference between the actual and theoretical rates of decrease in amplitude is doubtless due to a lack of isotropy, *i. e.*, to the imperfect elasticity and its variation from point to point. The effect of this heterogeneity must be to lengthen the period and wave-length and to retard the rate of decrease in amplitude as the wave moves on.

Efforts have been made to measure directly the rate of decrease in amplitude. Several difficulties have been experienced, one of which is the want of means of determining epifocal amplitudes in those quakes which are powerful enough to affect pendulums a quarter or half-way round the earth. These world-shakers in many cases originated at sea

and so far away from land that their vibrations which reach the shore are not perceptible to the unaided senses. The few that originate on land occur, with rare exceptions, in places where no measuring instrument is installed. And, finally, if a seismograph were installed near the epicentre of one of these terrible convulsions it would probably be wrecked before it could record a large amplitude. It only remains, then, to measure the amplitudes at different points along the path of a wave and at a considerable distance from the origin.

LONG-DISTANCE AMPLITUDES—MILNE

Catalogue No.	Origin	Place of Observation											
		Shide	Kew	Toronto	Victoria, B. C.	San Fernando, Spain	Bombay	Batavia	Mauritius	Mexico	Cape Town	Tokio	Philadelphia
250	Mexico	6/80	4.5/80	8/34	>17/30								6/34
344	Alaska	5/70		>17/20			0.5/105		.75/145			2.5/50	
345	"	5/70		>17/20			0.5/105		>17/145				
333	"	>17/70		>17/40	>17/20	20/77	>17/105		>17/145		10/165	1/50	
337	"			20/40		5/77	3/105			4/49	2/165		
338	"		>17/70	>17/40		20/77	9/105		4/145		10/165		
347	Ceram	2.5/121			2/105		1/62	10/22	1.5/73		0.5/105	5/47	
343	Smyrna	9/25					4/43		1/65		6/74	4/85	

The mark > indicates that the amplitude exceeded the upper figure, which was the largest the sensitised paper ribbon could record.

Professor Milne has given[1] the observed amplitudes at great distances in eight earthquakes occurring in 1900 or preceding years. Five of them originated in the sea west of Alaska and south of the Aleutian Islands, which has proved to be one of the most fertile breeding-grounds of world-shakers. The epicentres are so far from shore that the tremors are never felt on land and can be detected only by the pendulum. In the table on page 154 the amplitudes observed at twelve stations are given in millimetres placed above the line, while underneath is given the distance from the origin in degrees of arc of a great circle.

On this table Prof. Milne remarks:

> "Earthquakes like Nos. 343 and 347, from whatever may have been their amplitudes in the epifocal district, are reduced to an amplitude of 4 mm. after about 50° of travel. Larger earthquakes, like Nos. 337, 344, and 345, travelled 80° or 90° before their amplitudes sank to this quantity; whilst the largest of all, Nos. 333 and 338 show an amplitude of more than 4 mm. after travelling nearly half-way round the world. From an amplitude of 4 or 5 mm. the rate of decrease becomes less and less. For example, the amplitude of No. 337 between 77° and 105° falls from 5 mm. to 3 mm., or at the rate of. 07 mm. per degree; whilst from 105° to 165° the rate at which amplitude decreases has been .01 mm. per degree of travel."

The subject of periods and amplitudes has been most thoroughly investigated by Dr. F. Omori of the University of Tokio. For this purpose he used horizontal pendulums of his own design, so constructed as to meet the following requirements: 1st. The rate of motion of the ribbon receiving the record must be rapid enough to enable the observer

[1] *Brit. Assoc. Rep.*, 1900, p. 69.

to measure the periods of the different waves with accuracy. 2nd. The steady-point or bob of the pendulum must be brought so nearly to a state of neutral equilibrium that the period of its free oscillations shall be long enough to distinguish between the periods of the earth movements and the free period of the instrument itself. 3rd. The amount of friction between the parts of the instrument must be minimised. With several instruments constructed to meet these requirements and installed at the University of Tokio, observations were continued for eighteen months from June, 1898, to December, 1899. During this period 246 earthquakes were observed, all of which gave seismograms large enough to be accurately measured. These were divided up into nine groups, according to their points of origin, as follows:

Group I.	Distant earthquakes	95
Group II.	Originating off the eastern coast of Hokkaido (Yeso)	10
Group III.	Originating off the north-eastern coast of Honshiu (Main Island)	42
Group IV.	Originating off the coast of the provinces Hitochi and Iwaki	21
Group V.	Originating off the southern coast of Honshiu (Main Island)	3
Group VI.	Originating in Kiushiu or off its eastern coast	6
Group VII.	Originating in Central Japan	5
Group VIII.	Local earthquakes:	
	(*a*) Those observed at several places	38
	(*b*) Those observed in Tokio and one other place	10
	(*c*) Those observed only in Tokio	12
Group IX.	Miscellaneous origins	4

" Broadly speaking," says Dr. Omori, " the motion of an earthquake may be divided more or less definitely into three successive stages: the *preliminary tremor*, the *principal portion*, and the *end portion*, defined as follows:

"The *preliminary tremor* consists essentially of vibrations of small amplitude and of very short, or comparatively short, period. The *principal portion* denotes the most active part of an earthquake which follows the preliminary tremor and which consists of movements of larger amplitude. The *end portion* denotes the feeble finishing part of an earthquake, which follows the principal portion.

"In cases of large earthquakes proceeding from distant origins the preliminary tremor and the principal portion may each be further subdivided as follows. The preliminary tremor is made up of two distinct portions, which may be termed respectively the *first preliminary tremor* and the *second preliminary tremor.* The first preliminary tremor denotes the earlier portion, and the second preliminary tremor the later portion, the beginning of the latter being distinguished by a well-marked increase of the amplitude and, in many cases also by the appearance of slow undulations. In each of the two preliminary tremors the period remains on the whole constant, the amplitude also remains generally constant or, as often happens, rather greater at the commencement than toward the end.

"The principal portion is made up essentially of three successive parts, which may be termed respectively the *initial phase*, the *slow-period phase*, and the *quick-period phase.* The initial phase denotes the introductory part of the principal portion and consists of a few slow undulations. The slow-period phase follows the initial phase and consists of a number of slow undulations. These two phases are distinguished from one another by the difference of amplitude and period. The quick-period phase occurs toward the end of the principal portion and consists of comparatively quick waves. The period remains constant in each of the two last phases. In distant earthquakes the period remains essentially constant throughout the end portion. With earthquakes of near origins it may sometimes be different since there exist, in these cases, various kinds of waves, some of which may gradually disappear toward the end. The successive different stages of the earthquake motion are illustrated diagrammatically in the accompanying figure.

"In this diagram the different portions and phases are as follows:

a—Beginning of quake.
a c—Preliminary tremors.
a b—1st preliminary tremor.
b c—2nd preliminary tremor.
c d_3—Principal portion.
c d_1—Initial phase.
d_1 d_2—Slow-period phase.
d_2 d_3—Quick-period phase.
d_3 e—End portion."

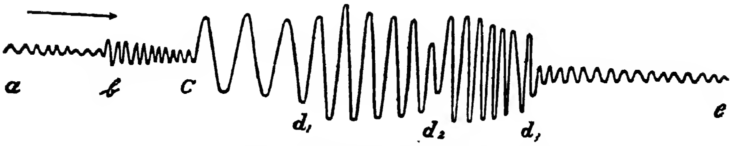


FIG. 51. Omori's Typical Long-Distance Seismogram.

Dr. Omori then gives a series of tables in the nine groups as recited above,[1] in which the details or elements of each of the 246 quakes are separately entered, with the date, hour, and minute of occurrence, the total duration of the quake and of each phase, the average period of each phase, and the maximum range of motion (double amplitude) in each portion. These tables would be unnecessarily long and voluminous for our purposes and are not given here. In a subsequent review with additional data he gives condensed statements in tabular form better suited to our convenience, which show the average period for each phase, the amplitudes and the durations of each phase and of the entire quake. These are reproduced. It should be noted that the wave-periods are averages in each phase. Dr. Omori finds that there are waves of many periods, though in each phase some two or three periods are much more frequent, while others are more or less uncommon. Thus in the first

[1] Publications of the Earthquake Investigation Committee in Foreign Languages, Nos. 11 and 13. Tokio, 1902 and 1903.

preliminary tremors of quakes of very different origin no less than five periods were identified, but much the greater part of them were of two periods, 4.6 seconds and 8.7 seconds respectively.

Mean Values of the Various Periods in the Successive Phases Deduced from 84 Quakes of Group I., *i. e.*, Quakes of very Distant Origin

1st P. T.		2d P. T.		Principal Portion						E. P.	
				I. Ph.		S. Ph.		O. Ph.			
No. of Quakes	Periods	No. of Quakes	Periods	No. of Quakes	Periods	No. of Quakes	Periods	No. of Quakes	Periods	No. of Quakes	Periods
14	1.04	2	1.03			3	.96	2	1.01		
26	4.6	2	5.1	3	3.3	2	4.1	5	5.7	1	4.8
28	8.7	14	8.5	3	8.4	6	8.6	32	9.3	32	9.6
8	15.	8	14.8			9	13.6	22	13.6	12	16.
						12	17.8	9	18.4		
2	20.8	3	20.3	5	22.9	9	22.3			1	22.
		3	25.5	6	27.6	10	25.9	1	25.8		
		1	32.6	4	34.3	2	34.3				
				4	43.3	1	45.4				
				2	54.0						

Mean Periods of 616 Quakes of Groups II. to IX., *i. e.*, Quakes Originating within 800 km. of Tokio or Nearer

1st and 2d P. T.		P. P.		E. P.		Mean	
No. of Quakes	Mean Period	No. of Quakes	Mean Period	No. of Quakes	Mean Period	No. of Quakes	Mean Period
	Sec.		Sec.		Sec.		Sec.
		4	.38			4	.38
		5	.61			5	.61
119	.92	165	.92	47	.98	331	.93
2	1.4	8	1.27			10	1.30
19	2.8	29	2.62	17	2.6	65	2.68
1	4.8	28	4.8	36	4.4	65	4 58
11	9.	38	7.7	34	8.4	82	8.17
1	12.1	26	11.4	1	11.	28	11.4
3	18.1	9	16.9			12	17.2
		7	21.8			7	21 8
1	24.	3	26.7			4	26.0
		3	29.7			3	29.7

The first table gives the periods for the several portions and phases of very distant quakes, *i. e.*, quakes whose origins are several thousand kilometres distant from Tokio, the place of record, and which therefore belong to the class of world-shakers. These all fall within Group I., recited above. The second table sets forth the periods observed at Tokio from quakes originating at moderate distances, either within the empire or in the neighbouring sea-bottom, and generally within a distance not exceeding about eight hundred kilometres, usually much less. The notation at the heads of the columns is as follows: 1st P. T., First preliminary tremor; 2nd P. T., Second preliminary tremor: P. P., Principal portion; I. Ph., Initial phase of principal portion; S. Ph., Slow phase of principal portion; Q. Ph., Quick phase of principal portion; E. P., End portion. In the second table the first and second preliminary tremors are not separated, as this is seldom practicable in any quakes except those of distant origin.

Dr. Omori is of the opinion that there is a tendency of these periods to fall into a comparatively small number of definite type-periods which preserve an approximate constancy. Thus in the preliminary tremors periods of about 4.8 sec. and 8.6 sec. largely predominate and they repeat themselves in the end portions of the quakes. In the principal portion there are several periods which most frequently recur. He suggests that this may be explained on the assumption that different portions of the earth's crust may have particular periods of free oscillation. This he thinks fits very well in the vibrations of the end portion, since the duration of an earthquake is long and the ground

must be performing wave-motion for a considerable time after the original impetus which caused the seismic disturbance has ceased.

One notable inference he draws is that these periods are not dependent upon the distance from the origin but are dependent rather upon the free period of the ground itself. Waves of quick period, however, are soon dissipated and are not of sufficient power to affect far distant instruments because of their absorption on the way. Long periods are not infrequent in quakes of near origin, and short periods also occur in the principal portions of quakes of remote origin. This view is not wholly in accordance with the general opinion that periods are generally longer as the distance of origin increases, though it is not pretended that the period is simply proportional to distance. Long periods of eight, sixteen, twenty-five, forty seconds, or more, do not appear to have been recorded in epicentral or meizoseismic tracts.

On the other hand, he finds that the duration of the quake and of its constituent portions increases in a simple ratio with the distance and also increases with the energy of the quake. Some of the distant disturbances lasted from three to four hours, and durations much greater still have been recorded elsewhere. This increased duration of the quake as the distance from the origin increases seems to raise a question in connection with Dr. Omori's view that the wave periods are not dependent upon the distance from the origin. For if the duration of the quake as well as the duration of its constituent portions and phases increases with the distance, then, since the quake and its constituent

parts are collectively a series of continuous wave periods, either the average wave period must increase with the distance, or the number of successive waves must increase. The latter alternative seems much less probable than the former.

CHAPTER IX

INTENSITY

Complexity of the Intensity Idea—Brief Illustration of the Motion of a Vibration Particle—The Acceleration of the Earth Particle—Variable Meanings Attached to the Idea of Intensity—Acceleration the Ordinary Measure of Intensity—It Is a Relation between Amplitude and Period—These Elements of Motion can be Accurately Measured Only by Means of the Seismograph—Estimates by Effects upon the Senses or Movable Objects—Usually the Only Ones Possible—The Rossi-Forel Scale — The Mercalli Scale — The Utility of these Scales — Checks upon Gross Errors of Estimate—Isoseismals—Estimates of Focal Depths—Holden's Attempt to Find Mechanical Equivalents of the Sensual Estimates by the Rossi-Forel and Mercalli Scales—Omori's Extension of the Estimates with Mechanical Equivalents Much Above the Highest of the Rossi-Forel

ONE of the most important quantities with which the investigator of seismic phenomena has to deal is usually termed the *intensity*. It is also the most perplexing on account of the different meanings given it. For what is meant by intensity? Obviously some earthquakes are more vigorous, more forcible, *i. e.*, more "intense" than others. Obviously, too, the same earthquake is more forcible, *i. e.*, more intense, in some places than in others,—is more intense at or near the epicentrum than far away from it. But in what does this intensity consist? Is it a simple or a complex idea? In some way it must consist of some form or aspect of energy, and according to present physical concepts

energy must be resolvable into terms of force, mass, and motion. And motion itself is further resolved into terms of mass, time, and space.

When we see a house with shattered walls, its front or side thrown out, its chimneys gone, and a few miles away a house which has only lost a little plastering or a chimney pot, we conclude that the first was more violently shaken than the second. But how much more violently? What was the intensity of the force in either case? In what units may that intensity be expressed and what ratio do the two intensities bear to each other? This is an example of the general question to which we now address ourselves, and to perceive its meaning it is necessary to look in some detail, though as briefly as possible, at the mechanism of wave-motion in solids.

Imagine the transmitting medium divided into a series of concentric spherical shells, the common centre of which is the point at which a disturbance or shock originates. A series of concentric spherical shells is not, indeed, strictly correct, but at a considerable distance from the origin it is sufficiently approximate for our purposes. As the disturbance is propagated outwards every particle is displaced elastically and returned to its original position. When the disturbance reaches the inner surface of a given shell the particles constituting that surface begin their excursions. The thickness of the shell may be so chosen that when the disturbance reaches its outer surface the particles on the inner surface have completed their excursions and returned to their original positions; the particles on the outer surface are just beginning to move; while particles of intermediate

position are in various "phases" of the movement. This idea is equally applicable whether the displacement be normal or transverse,—whether it be the propagation of elastic compression of volume, or of a mere twist or distortion.

The thickness of the supposed shell marks the distance between the points or surface where motion begins, and the surface where, at the same instant, the motion ends, and this thickness is termed the *wave-length*. The distance which the particle moves, or is displaced, is termed the *amplitude*. The total energy of a single impulse is conceived of as being at any moment of time contained between the outer and inner surfaces of this spherical shell. The shell itself is expanding outwards at a rate depending wholly upon the elasticity and density of the medium. Under conditions which are conceived of as ideally perfect, *i. e.*, where the elasticity and density of the medium are strictly uniform and homogeneous, the thickness of the shell, or wave-length, remains uniform, but the amplitude decreases in the same proportion as the radius of the shell increases; while the period, or time of its oscillation, remains constant. But its excursion grows less in the same ratio as the distance from the origin increases. The mean velocity of each particle, therefore, grows correspondingly less.

Since the energy is proportional to the square of the mean velocity of the particle, it follows that the vibrating energy of the particle is inversely proportional to the square of its distance from the origin. But the number of particles is directly proportional to the entire surface of the shell, which

in turn is directly proportional to the square of the distance from the origin. Thus, while each particle has less energy, the number of particles to which energy is imparted is increased in the same ratio. Hence the total energy is unchanged.

To impart motion to a particle, force must be applied. The amount of movement will depend upon the degree, or intensity, of the applied force, the time during which it acts, and the resistance which opposes its movement.

It is the intensity of this force with which we are just now concerned. In imagination we can conceive of the particle violently urged from a state of rest to a state of motion until the increasing elastic resistance to displacement stops the motion and brings the particle for an instant to rest, and then imposes upon it a motion in the opposite direction which returns it to its original place. The force which imparts motion is termed an accelerating force, and the amount of motion which it can impress upon a unit of mass in unit time is termed the acceleration. The standard of comparison is the amount of acceleration which the earth's gravitation can impress upon a gramme of falling matter in a second of time. In an earthquake the force of whose intensity we seek to gain some estimate is the force which produces the "*acceleration of an earth particle*," as it is termed.

It is important to note that acceleration is not velocity of motion but *change* in the velocity of motion. It may be positive or negative, *i. e.*, it may increase or diminish the speed, so that retardation is negative acceleration. Evidently the greater the amount of change in the motion of a

body, the greater the intensity of the force required to produce it. The vigour or intensity of a seismic vibration may, therefore, be considered as proportional to the *rapidity of the change* of velocity of the moving particles; or, in current phrase, proportional to the "acceleration of the earth particle."

The notion of acceleration as a change of motion may also be extended to include changes of direction, as well as changes of speed. In the shaking of the earthquake the earth particle is constantly changing the direction of its motion, and in a most irregular way. These changes of direction require the application of force, and the rapidity of the change is a measure of the intensity of that force.

This, then, is the force we are seeking, viz., the force that accelerates. It is the force which he who experiences the quake is conscious of. It is the force which moves objects upon the ground, which breaks, overthrows, and destroys. It is the force which causes the earth and everything upon it to tremble, vibrate, quiver, and rock. It is the force which does the work of the earthquake, becoming at once the measure of its energy and intensity.

It seems desirable to give mathematical expression to these various quantities, especially as some discussion has arisen concerning them. In doing so it becomes necessary to treat the vibration as a simple, harmonic motion; and though the motion is anything but simple, we shall incur no vital error in doing so. Let a be the amplitude, t the period, v the velocity of the earth particle, and I the intensity. Then, obviously, since velocity is space divided by time,

$v = \frac{2 \pi a}{t}$; and the acceleration $= \frac{v^2}{a} = \frac{4 \pi^2 a}{t^2}$.

In reducing the term intensity to mathematical expression it is necessary to have a precise understanding of the sense in which that term is used. On this point Professor Mendenhall very pertinently remarks (Am. Assoc. Adv. of Science, 1888):

"It has long been customary to speak of the intensity of an earthquake without any special effort to give the word an exact meaning. Generally it is applied to the destructiveness of the disturbance on the earth's surface, and sometimes to the magnitude of the subterranean cause of the same. But modern seismology proposes to measure the intensity of an earthquake, and to express its value numerically. It is worth while, therefore, to inquire in what sense the term may be used with precision and what may be accepted as its mathematical equivalent. Evidently it may mean, and in fact it has been made by different writers to mean, the measure of the surface of destruction, the energy per unit area of wave front; the rate at which energy is transmitted across unit area of a plane parallel to the wave front; and the total energy expended in the production of the original disturbance."

Remarking here that "energy per unit area of wave front" is merely a condensed expression for "rate at which energy is transmitted across unit area of a plane parallel to the wave front," the following is the mathematical expression:

$$I = \frac{2 \pi^2 a^2 V D}{t^2}$$

in which V=speed of wave transmission and D the density of the medium, both being for any particular wave regarded as constants.

The amplitude a and the period t are the quantities whose

values it is necessary to seek, and which can be found with definiteness and accuracy only by means of instrumental measurement. The seismograph is such an instrument, and its purpose is to give a measurable representation of the motion of the earth particle during an earthquake. It gives the period, the amplitude, and the general configuration of the motion in the three co-ordinate directions from which it becomes possible, under favourable conditions, to compute the acceleration at any moment, and, therefore, the intensity of the accelerating force.

But a seismograph is a very delicate, costly, and complicated instrument. Its instalment requires great care and skill, its attendance is expensive, and a severe draught upon the patience of the observer. The probability of securing from it a valuable record of an earthquake is usually very small. Instruments adapted to light or moderate quivers are unsuited to more forcible ones, and a severe or destructive shake is apt to wreck the entire instalment. Observed data, in order to be most useful for seismic study and analysis, ought to be numerous and well distributed over the affected areas. But seismographic observations must, from their costly, complicated, and exacting nature, be very few and far between. It therefore becomes necessary to have recourse to other means for securing the desired data

After a series of severe shocks has passed, it is in almost all localities certain that no seismograph has been installed to register them, and the only resource is in making rough mental estimates of the relative force of the shocks based upon the injuries or displacements which they have occasioned. At first thoughts it might seem as if this would be

tantamount to the vaguest kind of guesswork, and would be sure to lead to the grossest errors, so far from the truth as to be not only valueless but positively misleading.

If there were no means of checking gross errors in such estimates these apprehensions would be well founded. But such checks are readily applied, and means are at hand to prevent the errors from becoming so excessive as to make the results either valueless or misleading. There is one consideration which relieves the method of much of the distrust which might otherwise be justly attached to it. For many of the most important inquiries involving intensity estimates, we only require relative, and not absolute, intensities. We need to know only whether the intensity at one point was greater than at another and less than at a third point. There are, of course, some problems in which it is essential to have a very approximate estimate of the force in absolute terms of the C. G. S.[1] scale. But for these we have no resource except instrumental measurements, and must resign ourselves to the necessity of being content for the present with a very few of them.

To meet the necessity of some estimate of intensities, Professor di Rossi, of Rome, proposed some thirty years ago to employ a simple numerical scale to express the degrees of intensity inferred from the ordinary effects witnessed in an earthquake. In consultation with Professor Forel of Geneva he formulated a scale of ten degrees of energy as indicated by the effects upon men, structures, movable objects, etc. In the general absence of any other means of forming an estimate of the degree of energy exer-

[1] Centimetre, gramme, second.

cised by the quaking at the earth's surface, this scale was accepted everywhere with general favour, and has been found to be of very great utility.

ROSSI-FOREL SCALE

I. *Microseismic shock:* recorded by a single seismograph or by seismographs of the same model, but not by several seismographs of different kinds; the shock felt by an experienced observer.

II. *Extremely feeble shock:* recorded by several seismographs of different kinds; felt by a small number of persons at rest.

III. *Very feeble shock:* felt by several persons at rest; strong enough for the direction or duration to be appreciable.

IV. *Feeble shock:* felt by persons in motion; disturbance of movable objects, doors, windows; cracking of ceilings.

V. *Shock of moderate intensity:* felt generally by every one; disturbance, furniture, beds, etc. ringing of some bells.[1]

VI. *Fairly strong shock:* general awakening of those asleep, general ringing of bells; oscillation of chandeliers; stopping of clocks; visible agitation of trees and shrubs; some startled persons leave their dwellings.

VII. *Strong shock:* overthrow of movable objects, fall of plaster; ringing of church bells; general panic, without damage to buildings.

VIII. *Very strong shock:* fall of chimneys, cracks in the walls of buildings.

[1] House bells are evidently meant, such as were common before electric bells came into use.

IX. *Extremely strong shock:* partial or total destruction of some buildings.

X. *Shock of extreme intensity:* great disaster, ruins, disturbance of the strata, fissures in the ground, rock-falls from mountains.

There have been from time to time a considerable number of scales proposed by different investigators, all of which involve the same principle, though varying slightly in the specifications of the several degrees of intensity. These have been collated and published by Dr. Charles Davison (*Phil. Mag.*, 1900, vol. 50.) Some, in fact most of them, use only five or six degrees of intensity, their authors apparently believing that the use of ten degrees is an overrefinement with data of such a rough and uncertain character. The Japanese scale given by Dr. Davison has only three degrees, "slight," "weak," and "strong." It is by no means the only scale used in Japan, however. One of these scales, known as the Mercalli, is used in Italy and is published on the cover of each issue of the *Bolletino della Societa Sismica Italiana.* Owing to the importance attached to it in that country it is here given. It contains ten degrees, and the difference between it and the Rossi-Forel is not great, the chief distinction of practical value being such that the Mercalli scale seems better adapted to close gradations in the more powerful quakes, while the Rossi-Forel is better adapted to refinement in the lighter ones.

MERCALLI SCALE OF INTENSITY

I. *Instrumental shock, i. e.,* noted by seismic instruments only.

II. *Very slight:* felt only by a few persons in conditions of perfect quiet, especially on the upper floors of houses, or by only sensitive and nervous persons.

III. *Slight:* felt by several persons, but by few relatively to the number of inhabitants in a given place; said by them to have been "hardly felt," without causing any alarm, and in general without their being sensible that it was an earthquake until it was known that others had also felt it.

IV. *Sensible, or moderate:* not felt generally, but felt by many persons indoors, though by few on the ground floor, without causing any alarm, but with shaking of fastenings, crystals, creaking of floors, and slight oscillation of suspended objects.

V. *Rather strong:* felt generally indoors, but by few outside, with waking of those asleep, with alarm of some persons, rattling of doors, ringing of house bells, rather large oscillation of suspended objects, stopping of clocks.

VI. *Strong:* felt by every one indoors, and by many with alarm and flight into the open air; fall of objects in houses, fall of plaster, with some slight cracks in badly built houses.

VII. *Very strong:* felt with general alarm and flight from houses, sensible also out of doors; ringing of church bells, fall of chimney-pots and tiles; cracks in numerous buildings, but generally slight.

VIII. *Ruinous:* felt with great alarm, partial ruin of some houses, and frequent and considerable cracks in others; without loss of life, or with only a few cases of personal injury.

IX. *Disastrous:* with complete or nearly complete ruin of some houses and serious cracks in others, rendering them

uninhabitable; a few lives lost in different parts of populous places.

X. *Very disastrous:* with ruin of many buildings and great loss of life, cracks in the ground, landslips from mountains, etc.

At first sight such a scale seems to be not only liable to the greatest uncertainty of estimate, but so indefinite as to be misleading. It is apparently wanting in any definite quantitative relation whatever between the degrees of intensity which it proposes to formulate. What it really expresses is a rough estimate of the wall-cracking power, the window-rattling power, the man-scaring power, of seismic vibrations. What ratios can there be between the amounts of energy required to crack walls, to shatter chimneys, to dislodge plastering, or to break up a negro prayer-meeting? It would be rare to find a skilled observer of such phenomena who had ever experienced an earthquake, and the reports would be made by persons differing widely in temperament. At best it might seem as though they could not show much more than the bare fact that one area was shaken more forcibly than another without affording any indications of *how much* more forcibly. And this "how much" is the very thing which it is important to know. We shall soon perceive, however, that these rough intensity estimates have more value than their seemingly indefinite character might lead us to expect.

Isoseismals

It has been the custom of investigators in times past to collate reports of earthquakes from individuals who have experienced the shocks or tremors, and to arrange them

upon the map of the shaken district with estimates of the intensity, conformably to the Rossi-Forel scale, in each case; then to draw lines through the points of equal estimated intensity. These lines are termed isoseismals. They always form closed curves, or circuits, around the epicentral point which is usually not very far from the centre of the area which the curve incloses. Each degree of the scale, so far as observed, will, of course, have its own closed curve, and the curves of lower intensity will be outside those of higher intensity. They always show that as we recede from the central (*i. e.*, epicentral) point the intensity diminishes, and thus in a general way conforms to the theory. Let us look now at the theory.

For convenience we assume the earth to be a homogeneous elastic solid. We are to find the typical law in accordance with which the intensity should vary along a straight line upon the surface of the ground, passing through the epicentrum. In Fig. 52 let O be the origin, or focal point, of the vibrations situated at the depth E beneath the epicentral point E. Let the intensity at the epicentre be represented by a.

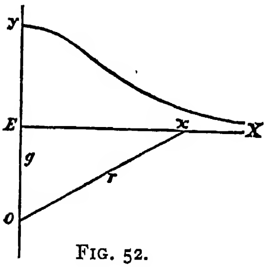


FIG. 52.

Since the intensity is inversely proportional to the square of the distance, the intensity at the epicentrum would be a. Take any other point on the surface, for example x, and connect it with O by the vector $Ox = r$. The intensity at x will be $\frac{a}{r^2}$. If we denote the intensity at any point on the surface by y we have,

$$y = \frac{a}{r^2} = \frac{a}{q^2 + x^2}.$$

This equation expresses a curve which may serve as a graphic representation of the way in which the intensity varies (under ideally perfect conditions) along a line radiating from the epicentrum. The variable y is here to be regarded as representing the intensity at any distance, x, from the epicentrum; the intensity at the epicentrum being $\frac{a}{q^2}$.

At a distance from the epicentrum equal to the depth of the assumed origin of the wave, the intensity, according to the above equation, would be only one-half that at the epicentrum. At a distance equal to twice the depth of origin, the intensity would be only one-fifth, and, at three times the depth, only one-tenth that at the epicentrum. These results are independent of the total energy of the earthquake, and are independent of the depth of its centrum. They are proportions which are true of all quakes, whether great or small, whether originating near the surface or at profound depths.

The records of the seismograph assure us that an earthquake which is forcible enough to be destructive at its epicentrum will be strongly felt at a distance very much greater than three times the focal depth. For such quakes, therefore, we should need a scale whose lowest intensity would be less than the thousandth part of its highest. Indeed it is probable that in quakes of the most forcible order, *i. e.*, No. 10 of the Rossi-Forel scale, their tremors may be quite sensible at distances where the intensity has fallen to one ten-thousandth of the intensity at the epicentrum. The roughly estimated depth of the Charleston earthquake was

PLATE VI


ISOSEISMALS OF THE
CHARLESTON EARTHQUAKE
ROSSI-FOREL SCALE
Scale of Miles.

about 12 ± 2 miles, and its tremors were noted at distances of nearly a thousand miles, or about eighty times the estimated depth. The intensities of this quake, estimated in terms of the Rossi-Forel scale, ranged from ten at the epicentrum to two at the distance of nearly a thousand miles. It is, therefore, well within probable limits to infer that the highest intensity was more than six thousand times greater than the observed lowest.

In view of the requirements of scientific methods the question arises whether it is possible to find some mechanical equivalents of the several degrees of intensity of this scale, which are capable of expression in algebraic terms and of treatment by ordinary algebraic processes. If so, the value of the scale will be much increased. This problem was undertaken in 1888 by Prof. E. S. Holden, then Director of the Lick Observatory, and the results he obtained, so far as they go, are highly interesting. He says:

"Referring to the Rossi-Forel scale we find that degrees 1, 2, 3, correspond to the feelings of the observer—to his sensations. The rest of the scale (4 to 10) refers chiefly to the effects of the shock in producing motion upon inanimate matter. The problem is to get some kind of common unit of a mechanical sort and to express the various degrees of the scale in terms of this unit. There is no question as to what unit to employ. The researches of the Japanese seismologists have abundantly shown that the destruction of buildings, etc., is proportional to the acceleration produced by the shock itself in a mass connected with the earth's surface. It would be logical to express (the intensity) in fractions of the acceleration due to gravity, *i. e.*, 9810 mm. per second. As these fractions are usually small, it is convenient to give the values of the intensity in terms of millimetres per second.

"The observations of Ewing, Milne, and Sekiya on Japanese

earthquakes give for each shock the amplitude and pe
which the velocity and acceleration can be computed.
quently a description of the effects of the shock on
etc., is given by them, which description is often suffic
nute to justify the characterisation of the shock by one
grees of the Rossi-Forel scale.

"I have carefully examined all the writings of these
named accessible to me, and after rejecting all doubtfu
have found twenty shocks, ranging in intensity from
which the amplitudes and periods were determined
ments, and in which I could assign the Rossi-Forel inte
confidence. The following table is the result:

Equivalents of the degree of intensity of earthquake shocks on the
scale in terms of the acceleration due to the velocity of the sho

Degree R.-F. Scale			Intensity		Diff.
1	corresponds	to	20 mm.	per sec.	—
2	"		40	"	20
3	"		60	"	20
4	"		80	"	20
5	"		110	"	30
6	"		150	"	40
7	"		300	"	150
8	"		500	"	200
9	"		1200	"	700

These results of Professor Holden are in the rig
tion, and it is desirable that they should be follow
more comparisons of the same kind. Twenty case
few for the reduction of such uncertain quantities
derived from mere eye-estimates to an absolute sca

These results of Professor Holden have been
mented by later investigations by Professor Omo
University of Tokio, who has made a long and caref
of experiments to ascertain by direct measuremen
celerations necessary to overturn or even fracture
and piers of various dimensions. The overturni

column by earthquake motion is a complex phenomenon, incapable of exact solution by a simple formula. But the problem can be simplified without losing its approximative character by the use of the formula $a = gx/y$, in which a is the acceleration, x the semi-diameter of the base, and y the height of the centre of gravity. Omori's experiments showed that this theoretical value of a differed in the mean of many cases from the experimental value only in the ratio of 1.07/1.

From these results he proceeded to construct an absolute scale of intensity founded upon the relations between the maximum accelerations and the damage produced in the great Mino-Owari earthquake. The scale applies, however, only to intensities above those classed as No. 6, in the Rossi-Forel scale. The lowest intensity, No. 1, of this absolute scale corresponds well with No. 7 R. F. No. 2 is considerably above Holden's evaluation of No. 8 R. F., while No. 3 coincides with the estimate of No. 9 R. F. Holden's evaluations end with No. 9 R. F., but Omori carries them on to three degrees higher than No. 10 R. F.

OMORI'S ABSOLUTE SCALE OF DESTRUCTIVE EARTHQUAKES[1]

"No. 1. Maximum acceleration = 300 mm. per sec. per sec. The motion is sufficiently strong that people generally run out of doors. Brick walls of bad construction are slightly cracked; plasters of some old *dozo* (godowns) shaken down; furniture overthrown; wooden houses so much shaken that cracking noises are produced; trees visibly shaken; waters in ponds rendered slightly turbid in consequence of the disturbance of the mud;

[1] Publications of the Earthquake Investigation Committee in Foreign Languages.

pendulum clocks stopped; a few factory chimneys of very bad construction damaged.

"No. 2. Maximum acceleration = 900 mm. per sec. per sec. Walls in Japanese houses are cracked; old wooden houses thrown slightly out of the vertical; tombstones and stone-lanterns of bad construction overturned, etc. In a few cases changes are produced in hot springs and mineral waters. Ordinary factory chimneys are not damaged.

"No. 3. Maximum acceleration = 1200 mm. per sec. per sec. About one factory chimney in every four is damaged; brick houses of bad construction partially or totally destroyed; a few old wooden dwelling-houses and warehouses totally destroyed; wooden bridges slightly damaged; some tombstones and stone-lanterns overturned; shoji (Japanese paper-covered sliding doors) broken; roof tiles of wooden houses disturbed; some rock fragments thrown down from mountain sides.

"No. 4. Maximum acceleration = 2000 mm. per sec. per sec. All factory chimneys are broken; most of the ordinary brick buildings partially or totally destroyed; some wooden houses totally destroyed; wooden sliding doors and shoji mostly thrown out of the grooves; cracks two or three inches in width produced in low and soft grounds; embankments slightly damaged here and there; wooden bridges partially destroyed; ordinary stone lanterns overturned.

"No. 5. Maximum acceleration = 2500 mm. per sec. per sec. All ordinary brick houses are very severely damaged; about three per cent. of the wooden houses totally destroyed; a few *tera*, or Buddhist temples, thrown down; embankments severely damaged; railway lines slightly curved or contorted; ordinary tombstones overturned; *ishigaki*, or masonry walls, damaged here and there; cracks one or two feet in width produced along river-banks; waters in rivers and ditches thrown over the banks; wells mostly affected with changes in their waters; landslips produced.

"No. 6. Maximum acceleration = 4000 mm. per sec. per sec. Most of the *tera*, or Buddhist temples, are thrown down; fifty to eighty per cent. of the wooden houses totally destroyed; embankments shattered almost to pieces; roads made through paddy-

fields so much cracked and depressed as to stop the passage of wagons and horses; railway lines very much contorted; large iron bridges destroyed; wooden bridges partially or totally destroyed; tombstones of stable construction overturned; cracks a few feet in width formed in the ground, accompanied sometimes by the ejection of water and sand; earthenware buried in the ground mostly broken; low grounds, such as paddy-fields, very greatly convulsed, both horizontally and vertically; sometimes causing trees and vegetables to die; numerous landslips produced.

"No. 7. Maximum acceleration much above 4000 mm. per sec. per sec. All buildings except a very few wooden houses are totally destroyed; some houses, gates, etc., projected one to three feet; remarkable landslips produced, accompanied by faults and shears of the ground."

Professor Omori then compares the foregoing absolute scale with the Rossi-Forel as follows:

ABSOLUTE SCALE ACCELERATIONS mm.—sec.	INTENSITY Cent. Meteor. Obs. Japan	ROSSI-FOREL SCALE
	Slight	1 2
	Weak	3 4 5
1—300	Strong	6 7
2— 900 3—1200 4—2000 5—2500 6—4000 7>4000	Violent	— 8 9 10 — —

CHAPTER X

VARIATION OF SURFACE INTENSITY

General Law of Decrease of Intensity away from the Epicentrum—Methods of Locating the Epicentrum—Seebach's Proposition—Determination of Points of Maximum Decline of Intensity—Method of Computing Depth of Focus—Intensity Curves as Dependent upon Focal Depth—Earthquakes with Deep Foci Compared with those of Shallow Foci—Maximum Focal Depths—Probably Never Greater than Fifteen Miles

THE grouping of intensity observations and their assemblage in the form of isoseismals furnishes us with the means of making important comparisons of earthquakes with each other. It furnishes us a scale of proportions by which we can judge, even though roughly, still with a fair and useful degree of approximation, the relative values of several important factors. The most striking of these are the depth of the origin, or ideal centrum, and the total energy of the earthquake.

The equation of the intensity curve already given contains the means of computing the depth of origin in absolute terms, provided certain facts in the distribution of intensity can be identified on the ground after an earthquake. Two points are required for this computation. The first is the epicentrum. The position of this point is obviously the centre of the isoseismal curves, unless local accidents of the ground lead to manifestations of intensity which are not

proportional to the inverse square of the distance from the origin. In case of such accidents our inferences as to the epicentrum would be more or less vitiated, and we should have no means of correcting our error. That this anomaly sometimes occurs is unquestionable. But ordinarily it is not of sufficient magnitude to lead to any large error in determining the point. The cases where it is liable to be most misleading are those quakes where the intensity at the epicentrum is small and the tremors extend over a large region. Where the intensity is high, and the disturbed district is not of extraordinarily great extent, there is seldom any difficulty in determining the epicentrum with a good degree of precision.

In those quakes which are generated by long dislocations or faults, or in those which result, like the New Madrid quakes, from the sudden sinking of a large tract of ground, the idea of an epicentrum is hardly applicable. There is in such cases no centrum, properly speaking. The causal action is as much superficial as subterranean, and is so widely diffused, and its distribution is so irregular and formless that the notion of a central, and, therefore, epicentral point, becomes incongruous. On the other hand, this diffused and irregularly distributed origin may approximate more and more to the relation of a centrum and point of origin as the distance increases. But at present we are concerned with the near neighbourhood of the origin and must exclude from our analysis those quakes whose origins are not of the compact and well-centralised kind.

Several methods have been employed by various investigators for locating the epicentrum. Among them is that of

Seebach, who uses the coseismals, or lines drawn through the point at which the first impulse, or perhaps the chief impulse, is felt at the same instant of time. According to the theory these should, under perfect conditions of propagation, form circles around the epicentrum, as in the case of the isoseismals. By determining three points of a co-

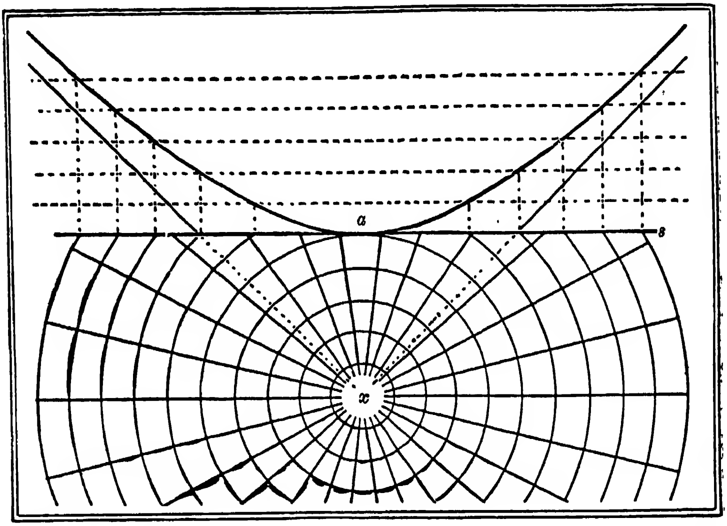


FIG. 53. Seebach's Method of Estimating Depth of Centrum by Means of Coseismals.

seismal curve the epicentrum follows at once. It is found at the intersection of the perpendiculars to two chords given by the three coseismal points, Fig. 53. The merits of this method are proportional to the accuracy with which the coseismal points are determined. These are time observations, liable not only to errors in noting the time, but also to much uncertainty as to what particular shock or oscilla-

tion is noted. Personal observation of the time is seldom well made. Instrumental observation is, of course, a question of the number of seismographs or "scopes" set at different places, and in most countries these are very few and far between. Moreover, instrumental time observations of seismic vibrations have a most unfortunate habit of breaking down or blundering in practice just when and where we most need them.

The coseismal method of determining the epicentrum, therefore, can seldom give a very accurate result. The same criticism might be passed upon the isoseismal method. But on the whole it is susceptible, under ordinary conditions, of greater accuracy and reliability than the coseismal. It has the highly important merit that it can be investigated deliberately after the quake is over, while the coseismal records must be caught at the instant the shock arrives or it is lost forever.

The determination of the second critical point of the intensity curve is much more difficult and uncertain than finding the epicentrum. This point may be described as the one at which the intensity diminishes most rapidly. By recurring to Fig. 54, it will be noted that between the epicentrum *a* and some point, say *d*, the curve is concave downwards. Beyond *d*, the curve becomes convex downwards. At *d*, therefore, there is a point of flexure, and the curve has then its maximum slope. And the degree of slope is a measure of the rate at which the intensity diminishes as we recede from the epicentrum. If we take the equation of the curve given above and differentiate it twice we have,—

$$\frac{d^2y}{dx^2} = \frac{8ax^2 - 2a(q^2 + x^2)}{(q^2 + x^2)^3}$$

At a point of flexure the value of the second differential coefficient becomes zero, and this value is satisfied in the above equation when

$$\pm x = \frac{q^2}{\sqrt{3}} \qquad y = \frac{3}{4}, \frac{a}{q^2}$$

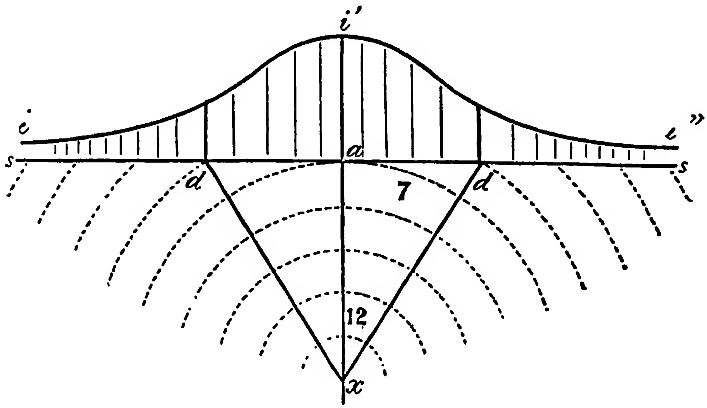


FIG. 54.

This value of y is found by substituting the particular value of x in the primitive equation of the intensity curve. It indicates that the intensity at the point of maximum rate of decline is three-fourths the intensity of the epicentre.

If q be considered as a radius and x as a tangent, then

$$x = \frac{q}{\sqrt{3}} = q \tan 30°$$

and the triangle ddx is equilateral. Or the angle ddx is 60°. In other words, the centrum or depth of the focus is found at a point where a line drawn from the point of maximum

decrease of intensity at an angle of 60° intersects the seismic vertical.

In the above value of x it is seen that the constant a (intensity at unit distance) has disappeared so that the distance of the point of inflection is independent of the original energy of the shock, or its intensity at the point of origin, and is dependent upon the depth alone. This is true of all earthquakes which have a distinctive or fairly well-defined centrum, whether they be powerful earthquakes or feeble. In all such quakes the distance from the epicentrum to the point where the decline of intensity is greatest is simply proportional to the depth of the focus *and to nothing else.* It is quite independent of the power of the shock. This property, equally applicable to all quakes, whether great or small, makes us independent of any absolute standard of measurement of intensity, and all that we require is to find the point (or circle around the epicentrum) where the intensity falls off most rapidly. The depth of focus is the radius of that circle multiplied by the square root of 3.

But it is not often that this point of maximum decline can be fixed. It can never be fixed with a high degree of certainty. It can seldom be fixed with a good degree of approximation, *i. e.*, where the probable error of the estimate is less than one-half the estimated depth. The causes of this difficulty are many. In the first place, very many earthquakes, in fact the great majority of them, have no well-defined centrum, but originate in disturbances affecting large, irregularly shaped tracts. The many vibrations of the earthquake issue from different points of that tract, and their individual centra may be a dozen miles apart, while

the theoretical point of maximum decline of intensity we are seeking may be only half that distance from the imaginary centre of the entire group of their origins.

Another difficulty arises from the fact that the decline of intensity is always gradual, though more rapid in some places than in others, and we have no means of measuring the gradations of intensity along a surface line except the Rossi-Forel scale, which at best is systematical guesswork rather than measurement. In other words, it may seem that accurate measurement, which is needed for the determination, is impossible through lack of any instrument for measuring. This difficulty, however, is checked in an important degree by the fact that the theoretical point must always be relatively near the epicentrum, and have an intensity which is three-fourths that which should occur at the epicentre. The observer who bears this fact in mind is, therefore, not very liable to locate the desired point much too far away from the epicentrum. If the intensity indications in the neighbourhood are abundant he ought to have no difficulty in recognising when the intensity has fallen to considerably less than one-half that of the epicentre.

We may illustrate this by an imaginary case. We have noted that at a distance from the epicentre equal to the focal depth the theoretical intensity should be one-half that at the epicentre. At a distance equal to twice the focal depth the intensity should be one-fifth that at the epicentre. At a distance of one-half the focal depth the intensity should be four-fifths that at the epicentre. The desired point, therefore, is located somewhere between the point where the intensity is one-fifth and the point where it is four-fifths

of the epicentral intensity. With abundant data of the ordinary kind it seems very improbable that the observer would locate it inside the higher intensity circle or outside the lower intensity circle. In other words, it is very improbable that he will underestimate his distance by one-half or overestimate it by double the true distance. Thus, under favourable conditions his probable error ought to be less than one-half the estimated distance.

The law expressed by the intensity curve can now be shown to be of great value and importance in the study of earthquakes even when our data are so rough and ill-defined as those which admit of no other treatment than the Rossi-Forel scale. It furnishes us a sense of proportion to which all our estimates of seismic quantities must conform, and it enables us to compare different earthquakes with each other. We may consider some comparisons, first in general terms, and then illustrate them by real occurrences.

Let us suppose a hundred pounds of dynamite to be exploded at a depth of a hundred feet in the ground (solid rock). It would produce quite a shock at the epicentre; but at the distance of half a mile the energy of the explosion would be only about the seven-hundredth part of the intensity manifested at the epicentre. Again: suppose ten thousand pounds of dynamite be exploded at the depth of a thousand feet. The effect at the epicentre would be the same as before, but at the distance of half a mile from the epicentre the intensity would be about one-seventh instead of one seven-hundredth the intensity at the epicentre. In these cases the intensity at the epicentre is assumed to vary inversely as the square of the depth and directly as the total

energy at the focus. In order that the epicentral intensity may be the same in both cases the total energy, or quantity of dynamite, must be proportional to the square of the depth.

Fig. 55 represents graphically a series of intensity curves

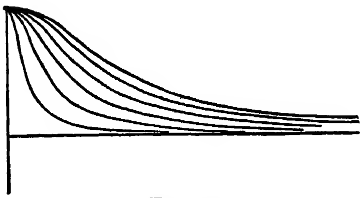


FIG. 55.

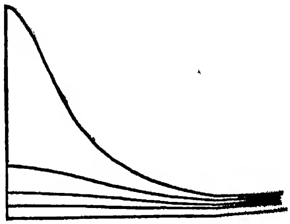


FIG. 56.

in which the depth varies in a simple ratio while the total energy varies in a duplicate ratio, so as to make the epicentral intensity the same in every case.

We may also suppose the total energy to be the same while the depth varies. The comparison of the intensity curves is shown in Fig. 56.

Finally, we may suppose the depth to be the same, while the total energy varies. The comparison will then be shown in Fig. 57. The difference in the last two cases, of course, consists in the fact that the intensity varies in the second case in a simple ratio with the energy, the depth being constant, while in the other case it varies with the square of the depth, the energy being constant.

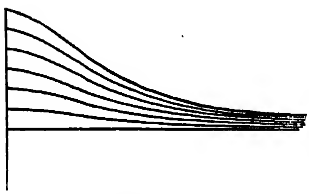


FIG. 57.

Thus the law of variation expressed by the intensity

curve becomes a normal scale or test by which we are enabled to keep our estimates of depth of focus from becoming wholly vague and random. By applying it we can avoid going very far astray from the truth, though we may not be assured of any high degree of accuracy. It establishes for us a sense of proportion which we cannot violate to any gross extreme without being made sensible that we are in error. Let us compare two earthquakes which offer some striking contrasts,—the Charleston event of 1886 and the Casamicciola disaster a few years preious.

At the epicentres the intensity seems to have been of the same order in both cases. The loss of life was much greater at Casamicciola because the visitors and residents of that gay resort were assembled in a large theatre, which fell with a crash at the first stroke, killing and maiming over two thousand people. At Summerville near Charleston, and close to the epicentre, the structures were wooden cottages which were rudely tumbled about and in a few cases thrown down, but the few villagers were more frightened than hurt. In Charleston itself, some sixteen or seventeen miles from the epicentre, where the estimated intensity was considerably less, structures of brick and stone were badly damaged, almost without exception, and many of them utterly wrecked. Less than a hundred lives were lost in the city.

Casamicciola is only about twenty-two miles from Naples. In that city the quake was felt only as a light or very moderate tremor, such as would be classed about No. 2 or 3 of the Rossi-Forel scale—certainly not higher than No. 4. The No. 3 isoseismal of the Charleston quake is drawn at distances from the epicentre averaging nearly six hundred

miles, and the No. 4 isoseismal at distances averaging about four hundred and fifty miles. Thus while the epicentral intensities seem to have been not very different the lighter tremors and oscillations of the Charleston quake were felt twenty or twenty-five times as far away as equally forcible tremors from the Casamicciola centre. From this we may infer that the centrum of the Charleston quake was twenty or twenty-five times deeper than that of Casamicciola, and that its total energy was from four hundred to six hundred times greater. The depth of the Charleston focus has been estimated roughly at about twelve miles, and if this be a fair approximation to the truth, the depth of the Casamicciola focus might be a little more than half a mile.

The latter earthquake would probably be regarded by many seismologists as a good type of those whose origin is clearly volcanic. The general result of observation throughout the world is that volcanic quakes are seldom of great energy, and have their foci at small or moderate depths. Though often highly destructive and disastrous their action is restricted to small areas, and the most destructive ones are not often felt at great distances. There have been a few cases, however, which seem to be exceptions to this inference. Perhaps the most notable one was the eruption of Krakatoa on the Straits of Sunda in 1883. This, however, was the most energetic volcanic explosion of the nineteenth century. It would seen, however, that the earth tremors, though well marked in Batavia and Buitenzorg and generally noticeable at distances considerably greater than is usual in volcanic eruptions, were not felt at half the distances which have been recorded in some great earthquakes not visibly

associated with volcanic action. The Krakatoa disturbance, however, was manifested thousands of miles away through the sea-wave which it caused.

The application of our intensity curve to the observed phenomena of earthquakes enables us to draw an important conclusion in this connection. It is that all earthquakes hitherto observed and recognised have their origins at very small depths as compared with the length of the earth's radius. All of them are only skin-deep. It is practically certain that none of them start from so great a depth as thirty miles. It is highly probable that none of them start from a depth so great as twenty miles. The argument in support of these assertions is furnished by the study of the intensity curve.

For suppose an earthquake shock with an energy as great as that of the Charleston event were to originate at a depth of a hundred miles. Its epicentral intensity would be about No. 7 of the Rossi-Forel scale. It would be alarming, it would shake down plastering, perhaps dislodge a few loose bricks from old chimneys. It might crack a few brick walls. People would run out of their houses, pictures would beat against the walls. But there would be no disaster—no wrecking of strong buildings, no distortion of the ground, no overthrow of pillars and monuments, no movement of houses on their foundations. But the vibrations would be felt at the distance of a thousand miles from the epicentre, where the tops of tall buildings would gently sway—chandeliers would slightly swing, very sensitive people might feel a nausea without suspecting the cause, and the delicate needles of magnetographs might show abnormal movements.

But such an earthquake has never been recorded. There have been shocks whose vibrations have been distinctly felt at a greater distance than a thousand miles, but their epicentral action has always been of the most appalling and destructive kind. There have been shocks whose epicentral intensity has been small, but it faded out quickly in the distance.

The point here brought forward does not propound the assertion that there is any antecedent reason why an earthquake may not originate deeper than twenty miles, or at any depth. That is a matter to be considered by itself. What is here urged is that in all recorded quakes, of which there is any reliable, detailed knowledge, there is none in which the ratio of epicentral to remote intensity would not conflict with the law of proportion set forth in the intensity curves, if it were assumed to originate at as great a depth as twenty miles. And that law is only a special case of the more general law that radiant or wave energies are invariably proportional to the square of the distance from the origin.

It is assumed, then, that there is a definite law of proportion governing the variation of intensity on the surface, and to that law the observed phenomena of all recorded quakes, so far as we have the means of judging, seem to conform. In that law the focal depth is a primary factor. So far as known, no shock has yet furnished evidence that this depth is more than twenty miles, and even that depth has not as yet been proven, or even so much as rendered probable, though it may be within the limits of possibility.

Thus far we have discussed the variation of intensity, as though the pulsations of the earthquake were transmitted

through a medium quite homogeneous in its properties of density and elasticity, and perfectly solid and continuous throughout its entire extent. We have still to reckon with the facts that the soils and subsoils of the earth are anything but homogeneous, and that near the surface, at least the stratified rocks are by no means continuous. That the defects of homogeneity and continuity must affect the transmission of the impulses in many ways is sufficiently obvious. But that it greatly affects the variation of intensity is not so clear. That it does so in a measure is not at all doubtful, but that it affects it sufficiently to vitiate seriously the general course of reasoning and deduction based upon it, is not apparent. The facts sustain the theory sufficiently to give the theory the character of a fairly satisfactory approximation.

In the discussion thus far it has been necessary to limit it to those quakes which have a centrum or something approaching that character, and to reserve remark upon those which originate in long faults or downthrows of large blocks of ground like the Chilian, New Madrid, Assam, and Sonora earthquakes. While this indefinite, diffuse, and often irregular configuration of the place of origin renders all our inferences more complex, it does not affect the fundamental considerations upon which our reasoning proceeds. These are still elastic waves originating from numberless points, more or less widely separated, sometimes simultaneous, sometimes in rapid succession, and all blending into a medley of vibrations like the sound-waves emanating from musketry and artillery in a battle. The quakes from the above causes are the most forcible and destructive, as well

as the most far-reaching of any that we know of. They are so because far greater amounts of energy are exercised in producing them than in the lighter quakes.

An opportunity was furnished in 1886 to compare the intensity of the Charleston quake at New Madrid with that of the New Madrid quake at Charleston. The distance separating the two places is about 575 miles. The Charleston quake at New Madrid was felt as a light tremor and oscillation on the ground, but as a long swaying motion in upper stories of buildings. It was noticed by only a few people and caused no alarm. The New Madrid quake was much more forcibly felt in Charleston, where it cracked walls, set the church bells ringing, and threw down plastering. Even in Boston, over twelve hundred miles distant, it was apparently forcible enough to attract general attention and tc cause considerable alarm.

CHAPTER XI

SPEED OF PROPAGATION OF EARTH-WAVES

Importance of this Question in Relation to the Problem of the Earth's Interior Condition—Difficulty of Ascertaining Wave Speeds—Especially at Small or Moderate Distances—Three Kinds of Waves, Each with its Own Speed—Wide Discordances of Results—Difference between True and Apparent Speed—Hopkins's View—Seebach's Method of Finding Depths of Origin from the Relation between True and Apparent Speed—August Schmidt's Proposition Concerning Effect of Variable Elasticity upon Wave Speeds and Wave Forms—Results Obtained over Short Distances—Speeds in the Charleston Quake—Speeds Obtained by Omori in Japan—Indian Quakes Discussed by Agamennone—Real Causes of the Difficulty of Securing the Proper Wave Data

NO branch of the subject of earthquakes has received more attention than the speed with which their vibrations are transmitted. A knowledge of this quantity has been deemed valuable in attempting the solution of some of the greatest questions of dynamical geology. For the rate of transmission is assumed to be dependent wholly upon the elasticity and density of the transmitting medium. The speed of propagation, then, ought to indicate to us something as to the elasticity and density, and, therefore, of the physical condition in general of those materials which form the interior of the earth. Nor do we seem to have at present any means of extending our inquiries experimentally into those regions. But an earthquake vibration which has

traversed a diameter or a long chord of the earth's interior may be supposed to have brought us some message from the mysterious realms it has crossed on its way, though the difficulty of translating it aright may be insuperable.

At first it might seem a very simple matter to time the passage of a shock at two points. The difference of time and the distance between the two points being known, the speed follows at once. But nature is not half so simple, and is seldom or never disposed to surrender her precious secrets so easily. Instead of timing a single, short, well-defined impulse, the observer is required to note a protracted series of them, whose beginning is more or less uncertain, whose increase is gradual up to a maximum, or even several maxima, and whose ending is through a progressive decline. The speed of transmission is always swift, several kilometres per second, and, unless the space interval between observing stations is considerable, a very small error in the observed time leads to a large error in the result. In the vast majority of earthquakes such errors are inevitable. It is only in the more powerful ones that long intervals of both distance and time are available for measurement, and these are rather uncommon.

Again, there are three distinct classes of waves whose speeds are different, being dependent upon different sets of physical constants, viz., the normal, the transverse, and the surface waves. Near the epicentre these are all more or less intermingled. As they spread out they separate, but even within two or three hundred miles of the origin the separation is so incomplete that they still overlap each other more or less, and it is not until they are two or three thousand

miles away from it that the separation is fully completed. By that time the amplitudes of the vibrations are so far reduced that only instruments of extreme sensitiveness can pick them up and properly record them. The normal and transverse waves also reach great distances by passing through the deeper portions of the earth's interior, the rays following paths which are possibly not chords, but curves due to a variation of the ratio of elasticity to density, and converge toward the earth's centre. The surface waves, true to their name, follow around the earth in arcs of great circles. By the time they reach a recording instrument four or five thousand miles away, they have travelled a greater distance from the origin than the normal and transverse waves.

Finally, the purely elastic waves, on their way through the depths, may perhaps cross regions of varying ratio between elasticity and density, and undergo changes of speed, probably increasing at first as their paths get nearer the earth's centre, and afterwards decreasing as they get farther away from it.

These few remarks may suffice to suggest the complexity of the subject. They may also explain why recent investigators have divided that subject into two distinct fields, one of which comprises speed measurements and estimates over short distances from the epicentre or origin, and the other, those which deal with speeds through long distances of several thousand miles from the origin. Both fields of inquiry have yielded instructive results, the latter or long-distance estimates being the more valuable of the two groups. But some knowledge of the results of

short-distance measurements is essential to a full understanding of the others.

The earliest attempts to measure the speed of earth-waves approached the subject in two ways. The first method consisted in producing artificial shocks by means of falling weights or gunpowder explosions, and measuring the time interval occupied by the tremor in passing from its origin to the point of observation. We have incidentally alluded, in Chapter VII, to some of these experiments by Abbot, Milne, etc., and have seen the wide discordances in the results, which ranged from one or two hundred metres per second to 2700. Such results must obviously yield more questions than answers.

The second method of approach consisted in collecting the best available time records showing the instant of the passage of an earthquake shock at several points—indeed, at as many points as possible. The difficulty of securing such time reports was great. It soon appeared that the speeds were far greater than had been supposed, and would have to be reckoned in kilometres per second rather than in hundreds of feet, and accurate time-observations would be necessary. But observations of requisite accuracy could rarely be expected except from astronomical observatories and from the few individuals or establishments whose functions required them to keep accurately regulated clocks. Another difficulty arose from the fact that an important earthquake, or one forcible enough to be felt at a distance of one or two hundred miles, was an ill-defined event in respect to time. It usually lasted a minute or more at the epicentre, and the duration of the shaking increased with

the distance. It was usually impossible to decide to what particular phase, part, or vibration a time record belonged. When the tremors had become enfeebled by distance it was presumed at first that there would be some one great oscillation which could be regarded as the shock *par excellence*, and be distinguishable wherever all others might fail. But though this might in a very few cases be true, it was generally not so.

It is not surprising, therefore, that the results varied considerably, ranging, in fact, from one and a half to six kilometres per second, and showing a tendency to cluster around an average of about 3000 or 3100 metres per second, as the mean rate of propagation of such waves as are ordinarily recorded upon seismographs within two or three hundred miles of an epicentre. Wide apart as the various estimates were, they were very much less discordant than those obtained in experimental shocks, and may be regarded as the first rough approach to true determinations. For a long time, however, so little was the complexity of the problem appreciated that some investigators were disposed to rely upon assumed rates of propagation to supply data for estimating the depth of origin of an earthquake.

Professor Hopkins, of Cambridge, had pointed out the distinction between true and apparent wave speed. In Fig. 53, let x be a compact, well-defined centrum. Let the plane SS represent a portion of the earth's surface, and let the curves on SS, as well as those on the co-ordinate vertical plane through $a\,x$ represent successive, equidistant positions of the spherical wave emanating from x. The true speed of propagation is the speed along the rays or vectors, $x\,b$,

xc, xf, etc. But the apparent speed is that which is observed along the line af, and it is evident that they are not the same. If ϕ represent the angle which the seismic vertical makes with any ray (say xc), and if v is the true speed and v the apparent speed, it is evident that $\mathrm{v} = v/\sin\phi$. At the epicentre, therefore, where $\phi = 0$, the apparent speed becomes infinite, while at a great distance, where ϕ approximates $90°$, v and v become sensibly equal. At a considerable distance, however, say two thousand or three thousand kilometres, the curvature of the earth enters as a modifying consideration.

Assuming the foregoing law of Hopkins, Prof. Karl von Seebach proposed a method for finding the depth of the centrum or origin of a wave. As we have just indicated, the depth, or seismic vertical, involves a relation between the true and apparent speeds. If we assume that the true speed is constant, then the relation becomes a function of the apparent speed and of the distance of the wave front from the epicentre, both of which Seebach assumed to be measurable with accuracy. By plotting the spaces traversed on the surface in equal times upon the X co-ordinates, and plotting the equal times on the Y co-ordinates, and drawing a line through the ends of the latter, the result is a rectangular hyperbola whose asymptote passes through the centrum. From the equation of this hyperbola the value of the seismic vertical is at once deducible.

Seebach's proposition, however, failed to yield any acceptable results, and the reasons are now plain. Its application required data whose accuracy must be great, and for reasons given at the beginning of this chapter, as well as for

others yet to be described, such data are quite unattainable. Moreover, Seebach had the same misconceptions as all his contemporaries about the nature of surface vibrations in an earthquake; misconceptions which even now have been only partly cleared up by improved instruments and methods of investigation.

Seebach's proposition involved the implied assumption either that the earth was homogeneous in respect to elasticity and density, or that the ratio of elasticity to density was sensibly constant.

Schmidt's assumption was that elasticity varied (and presumably increased) with the depth. In passing from a less to a more elastic medium the speed increased. The waves, therefore, instead of remaining spherical became elliptical.

A ray being always normal to the wave front, therefore, becomes a curve instead of a straight line as Seebach's proposition contemplated.

Schmidt's proposition is conditioned only upon the assumption that the true speed of the wave is always the same at the same depth, though in passing from one depth to another the variation may follow any law. The proposition should remain true even if the wave speed were to diminish with the depth, only in that case the rays would be concave towards the earth's centre, and only a few of them would reach the surface. But as we have good reason to believe that the speed of the transverse wave at least increases centrewards, the rays must be convex towards the centre.

As no means existed for measuring or estimating variations of elasticity and density within the earth, no numerical calculation of the amount of bending in the rays was

possible, and Schmidt's law furnished no additional means of applying surface speeds to measurements or estimates of focal depths. The proposition, however, has uses in other relations.

We may now proceed to examine some of the results of efforts to ascertain wave speeds from time observations over

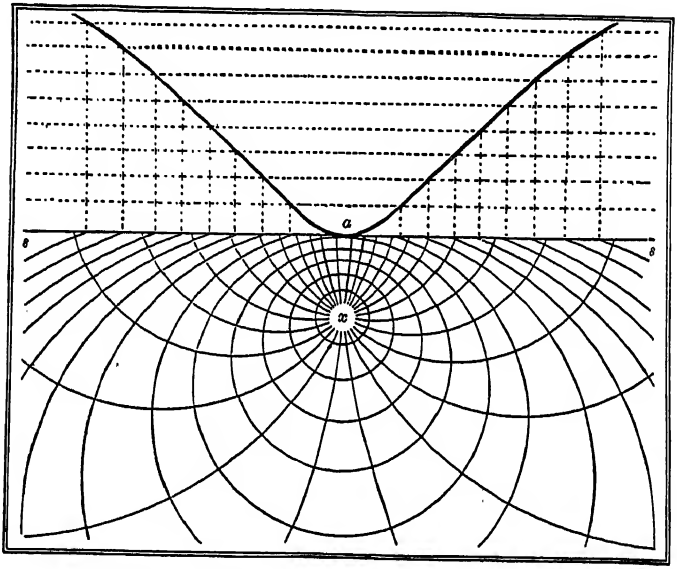


FIG. 58. Aug. Schmidt's Inference of Refraction.

paths of less than a thousand miles. Since the first difficulty to be met is the attainment of accurate time-data, it is obvious that the greater the length of path the less effect has a given error upon the result. Very powerful and far-reaching quakes, therefore, may be expected to furnish better data than weak ones. In a large region well-peopled and well-civilised, like Europe or North America, the num-

ber of observations is very liable to be proportional to the area, which in turn is proportional to the square of the distance at which the quake is distinctly perceptible. In Europe, however, nearly all of the very powerful quakes have their origin near the southern ends of the long peninsulas which extend into the Mediterranean (Andalusia, Calabria, Greece), or in the floor of the Mediterranean itself. Hence they affect far greater water areas than land areas, and the populations of the areas affected by them are about the least likely of any in Europe to make accurate time-observations.

Again, the generally diffused habit of keeping clocks and watches which are frequently corrected to conform to some definite standard of time has been the development of the last three or four decades. It began with the introduction and wide extensions of railways and telegraphs. The growth of industry and commerce, along with the increasing regularity of all communications, has brought with it the increasing necessity for punctuality in all occupations, and, therefore, the necessity of fairly well-regulated clocks and watches in every shop, factory, and office, and in almost every man's pocket.

The first great earthquake perceptible a thousand miles from its origin throughout an area one-half, at least, land, filled with a highly civilised people accustomed to keep fairly accurate time was the Charleston earthquake of August 31, 1886. This occurrence was investigated with as much thoroughness as practicable by the United States Geological Survey, and the most earnest efforts were made to secure all practicable time-data. The number of time

reports obtained, after rejecting those which on their face were plainly of no value, was 316. Of these, 130 were rejected after careful scrutiny. The remaining 186 were divided into four groups. So far as possible the endeavour was to establish in each observation the nearest practicable approach to the time when the tremors first became perceptible. The first group, which contained only five observations, gave the times of the beginning with an uncertainty not exceeding fifteen seconds. The second group, containing eleven observations, gave times of beginning with an uncertainty not exceeding half a minute. The third group comprised 125 observations without reference to the phase of the quake, and only requiring that the time be taken from well-kept clocks. The fourth group comprised stopped clocks[1] which had been regulated by standard time. The speeds computed from these groups were:

Group I.	5205 ± 168	metres	per	second
Group II.	5192 ± 236	"	"	"
Group III.	4848 ± 43	"	"	"
Group IV.	4245 ± 168	"	"	"

[1] It might be supposed that a stopped clock, if well regulated, would be the most reliable witness possible of the time. So it would be if it could be relied upon to stop at the beginning, or at any other determinable phase of the event. But investigation proved that the clocks which were stopped ceased their motion near the end of the quake. As we now know, the long waves of the quake are always the concluding part of it, and it is these long waves, apparently, which stop fine, accurate clocks with pendulums beating seconds. The times of the stopped clocks were in every case where comparison was possible considerably later than directly observed times reported by individuals, and the tardiness in stopping was nearly proportional to the distance from the epicentrum. As compared with the other groups, the stopped clocks furnish notably longer time-intervals, and, therefore, a slower speed. Applying an estimated correction for this discrepancy, the speed deduced from stopped clocks would be brought into agreement with those derived from the first two groups.

The speed thus inferred came as a surprise to most seismologists, as it very greatly exceeded any that had been credited before. But none had ever been sustained by so large and concordant a mass of trustworthy observations nor over such extensive areas. Since that time the speeds of a considerable number of quakes have been computed in different parts of the world over paths of less than a thousand miles. A very few have given as great or slightly greater speeds. But in the great majority the speeds have been considerably less.

In a paper by Prof. F. Omori,[1] of the University of Japan, the speeds of the waves from four of the greatest Japanese quakes between October 28, 1891, and October 22, 1894, inclusive, are discussed. The distances over which the speeds are computed range from 130 to 730 kilometres. The following are the results:

Earthquake with After-Shocks	No. of Shocks	Average Speed per Second
Mino Owari, Oct. 28, 1891......	17	2.1 km.
Noto, Dec. 9, 1892	2	2.1 km.
Hokkaido, March 22, 1894......	8	1.8 km.
Shonai, Oct. 22, 1894.......	1	2.0 km.

A series of powerful shocks originating at Zante in the Ionian archipelago, in January and February, 1893, were recorded on the seismographs of Italy and other European stations, and the speeds have been discussed by Dr. G. Agamennone.[2] In the accompanying diagram, Fig. 59, the

[1] F. Omori. "Sulla velocita di propagazione e sulla lunghezza delle unde." *Boll. Soc. Seismo. Ital.*, 1895.

[2] *Atti della Reale Accad. dei Lincei*, Seduta del 17, Dicem., 1893.

horizontal scale represents the distance named in kilometres, and the vertical scale indicates time intervals, the spaces be-

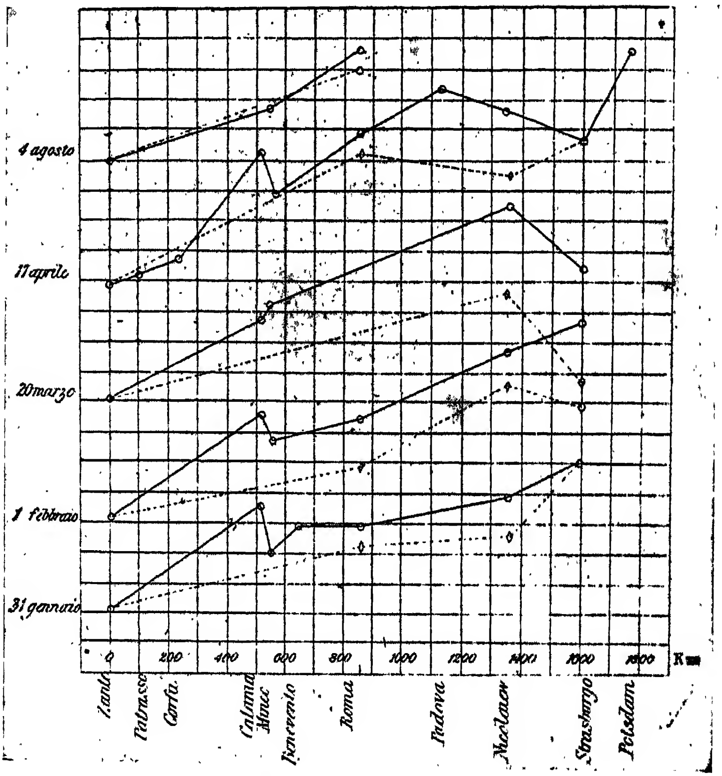


FIG. 59. Dr. Agamennone's Speed Diagram of Five Zante Earthquakes.

tween the horizontal lines corresponding to intervals of twenty seconds each.

It will be noted that the speed, as indicated by Dr. Agamennone's diagram, varies inversely as the inclination of the line to the horizontal axis. The less the inclination the

greater the speed. The following table indicates the resulting speeds under three different conditions. The first is computed by treating all accepted time-data to the process of least squares, irrespective of whether they refer to the beginning or to the maximum phase of the vibrations. The second uses data which refer to the maximum phase only. The third uses data referring to the beginning only.

SPEEDS OF ZANTE QUAKES IN 1893. AGAMENNONE

Date 1893	Speed Based on all Obs. Regardless of Phase		Speed of Maximum Phase		Speed of Initial Phase	
	No. of Stations	Km. per Sec.	No. of Stations	Km. per Sec.	No. of Stations	Km per Sec.
Jan. 31......	7	4.04	4	2.86	4	3.08
Feb. 1......	6	3.28	4	2.42	4	3.92
Mar. 20.....	5	2.33	3	2.82	3	7.79
Apr. 17......	10	2.34	5	2.23	5	2.64
Aug. 4	3	2.12	2	2.36	2	2.83
Mean.		2.34		2.43		3.08

The foregoing investigations may suffice to show the rather large discrepancies in the speeds of propagation over distances of less than a thousand miles. Many attempts have been made to explain them and to reconcile them with wave theory and with each other. But at every turn these efforts are confronted with several serious difficulties. One is the impracticability of separating the different kinds of waves (normal, transverse, and surface) and the impossibility of distinguishing them. Another is the different degrees of sensitiveness in the recording instruments, which vary from country to country in structure and mode of performance,

the result being that it is seldom possible to be confident of their time records. As Professor Milne very justly remarks[1]: "It does not seem likely that, until we are in possession of a series of records taken at long distances apart on the surface of the globe by means of instruments which are similar, which have sufficient sensibility to record preliminary tremors, and which record upon surfaces moving sufficiently quickly to allow of accurate time determinations, that our present knowledge will be greatly increased."

The question has often been raised whether the speed of propagation is constant. Theoretically it should be constant only when propagated through media in which the elasticity-density ratio is constant. That this ratio is a variable one is always taken for granted. This is equivalent to saying that constancy cannot be expected. It has, however, been inferred by many that the speed increases with the distance, and by a few that it diminishes with the distance. Both views, especially the first, can find support in the computed speeds of different quakes. Instances can be cited which appear to indicate in the first few hundred miles a decrease of speed, followed at greater distances by an increase of speed. But these cases really prove nothing. When we remember that the instruments vary much in delicacy and pick up a wave or wave-group early or late according to their sensitiveness, we may expect to find evidence supporting all kinds of views on this subject.

[1] *Rep. Brit. Assoc. for Advancement of Science*, 1895, p. 169.

CHAPTER XII

LONG-DISTANCE WAVE SPEEDS

Origin of Present Pendulum Observations—Von Rebeur-Paschwitz's Investigations of Minute Deflections of the Plumb-Line—Milne's Researches—Identification of Such Disturbances with Distant Earthquakes—Dr. Charles Davison's Observations—Von Rebeur-Paschwitz's Suggestion of Co-operative International Inquiry upon this Matter—Researches of the Italian Seismologists—Accumulation of Long-Distance Seismograms—Their Study by Prof. C. G. Knott—Results of the Study—Different Groups of Waves—Preliminary Tremors—The Following Phases—Dr. Oldham's Researches—Speed of Surface Waves—Speeds of Preliminary Tremors—Dr. Knott's Remarkable Results—Method of Determining the Origin of a Distant Earthquake

AS far back as 1883, Professor Milne, animated, perhaps, by the brilliant success he had even then attained in devising instruments for seismic investigation, ventured to predict (in substance) that the time would come when a great earthquake would be recognisable at any point on the terrestrial surface, even as far away from its origin as the antipodes. His meaning appears to have been that a forcible quake generates vibrations which permeate the whole earth within and without, and that its recognition at any point is only a question of suitable instruments. It would seem as if he had never lost sight of his prediction, which has been fulfilled sooner, perhaps, than he might have been justified in expecting.

This recognition has been gradually reached by the co-operative results of two lines of experimental inquiry, both involving a common problem, viz., the constancy or inconstancy of levels. Astronomers had long been aware of certain anomalies in the records of their instruments which were explainable only by minute changes in the levels of the piers on which they were installed. These suspected changes of level were usually temporary and were soon recovered, being, in fact, periodic. They could not be attributed, therefore, to the settling of their foundations, nor to any permanent changes of the ground beneath them. Some of them have been found to correspond with the daily oscillations of the barometer; others, of annual period, to the seasonal changes. The amounts of change seldom exceeded one or two seconds of arc, but the increasing standard of accuracy in astronomical work made this a very important matter. Systematic inquiry into these changes was, therefore, begun in England, France, and Germany as far back as 1865 by means of instruments of extreme sensitiveness. The work of the seismologists in Japan and Italy was well known, and the investigators in the interest of astronomical science were naturally on the watch to ascertain whether seismic disturbances might not be revealed in the analysis of the records obtained from their instruments for detecting minute fluctuations of level. Such disturbances were definitively recognised in 1889 by Dr. E. von Rebeur-Paschwitz at Potsdam and Wilhelmshaven.

Meantime investigations were in progress by Professor Milne in Japan, having for their purpose the analysis of the various classes of tremors which were represented in the

traces given by horizontal pendulums. He was led to the conclusion that the column on which the instruments rested was subject to periodic tilting by movements of the ground resembling the long, flat swells of the sea, originating in a distant storm. These tilts repeated themselves at intervals of two or three seconds, and the deflections of the pendulums were equivalent to tilts varying from one to four seconds of arc. There were also indicated other movements of two or three minutes in period, and these periods were remarkably regular. These traces were sent to Von Rebeur-Paschwitz, and the comparison left very little doubt that both were dealing with the same kind of phenomenon. It was suspected that these records were caused by far-distant earthquakes; but considerable time must elapse before the suspicion could be verified or dispelled by the accumulation of more evidence. And the evidence was at length forthcoming.

As the observations of Von Rebeur-Paschwitz were primarily for astronomical purposes he secured the co-operation of Professor Kortazzi in the Russian Naval Observatory at Nicolaieff, to whom he furnished one of his very delicate horizontal pendulums described in Chapter VI. Similar ones were set up at Potsdam, Wilhelmshaven, and Strassburg. It was soon shown that a very large proportion of the suspected disturbances were common to all four places. Indeed, as Von Rebeur-Paschwitz says:

"It is a comparatively rare occurrence that when an earthquake figure, however small, appears on one of the photographs, it is not equally visible on the other. It often happens that the curves are not sufficiently distinct, owing to variations in the

intensity and figure of the light-point, and faults in the paper, or where a general microseismic movement is more pronounced at one of the stations than at the other. In such cases small disturbances may at first escape detection, but are often found when notice is given from the other station."

It was also found that these disturbances followed a few minutes (less than half an hour) after some earthquake of very considerable force in a distant part of the world. Generally the time interval between a great quake and the appearance of the indications of it upon the photographic records of the pendulums bore definite proportions to the distance traversed by the impulses. But these definite proportions were found to hold only in one and the same earthquake recorded at different distances from the origin. As between different earthquakes it appeared that the more powerful ones were propagated a little more rapidly than the weaker. The general agreement in the character of the indications given by the various pendulums was a strongly corroborative feature.

In the columns of *Nature*, December 27, 1894, Dr. Charles Davison called attention to a publication of Dr. von Rebeur-Paschwitz, making public for the first time, so far as known, the diagrams obtained from the latter's pendulums. The accompanying Fig. 60 is reproduced from that article, and shows two photographic traces, one obtained at Strassburg, the other at Nicolaieff. The numerals on the lower line indicate hours of Greenwich mean time. It will be noted that both traces are interrupted by gaps which are due to several causes, the chief one being the wide oscillations of the light-point which often carried it off the band

of sensitised paper, and partly to the to-and-fro movement being of such amplitude and the light-point moving so rapidly across the paper that it failed to leave a trace. Wherever these gaps occur the mind must supply oscillations larger than those which are shown. The individual oscillations do not appear because of the very slow movement of the paper ribbon, viz., about half an inch per hour.

On that same evening and at about the same hours, Dr.

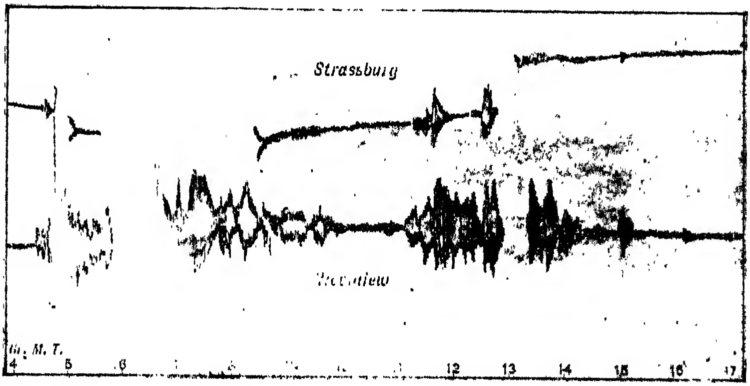


FIG. 60. Von Rebeur-Paschwitz's First Long-Distance Seismograms at Strassburg and Nicolaieff.

Davison, who was then conducting observations at Birmingham with the bifilar pendulum, saw its image moving slowly from side to side, indicating the passage of long, slow undulations of very small amplitude. This instrument, it will be remembered, is insensitive to rapid tremors, but extremely sensitive to slow tilts or variations of the vertical. Its indications began after 5.43 P.M., Greenwich, M. T., but the observer did not reach the instrument until 6.29, which corresponds to the principal gaps in the Strassburg and

Nicolaieff records, when the movements at both those places were so large that the records were lost. At 6.42 the image of the bifilar pendulum had come to rest. But at 6.46 the oscillations began again and continued with varying amplitude until 8.13. After that hour no further motion was noticed, though the observer watched for two hours and a half.

With the knowledge since gained it is now easy to fill up the gaps in these records and perceive their entire agreement. In this instance, however, they have not been identified with any earthquake of a high order of energy. Dr. von Rebeur-Paschwitz, after a study of all the observations, inclines to the view that the origin was far to the south-west of Strassburg, which would seem to indicate a submarine origin, as it would surely have been reported if its epicentre were anywhere upon the land.

From this time forward the observations of these disturbances increased gradually. It was seen that the phenomena were of world-wide extent, needing the co-operation of many observers stationed in many parts of the world. Upon the initiative of Dr. von Rebeur-Paschwitz, warmly seconded by Professor Milne, observatories were provided with suitable instruments in numerous countries, and something like a general spirit of co-operation prevailed among them. The British Association for the Advancement of Science, yielding to the earnest spirit and enthusiasm of Professor Milne, lent its assistance and prestige to the organisation of observatories throughout the British Empire. The Royal Society successfully approached the Government for the slight financial aid needed for the work. In the

British Association Report for 1898, and for several subsequent years, Professor Milne assembles and discusses a large number of these distant observations and reproduces the traces of a considerable portion of them. They form collectively a contribution to seismology whose value and interest cannot be overestimated.

Meantime, the Italian seismologists were making inquiry into the same phenomena with a different class of instruments, and their results may be regarded as supplementing in a very desirable manner those obtained from the horizontal pendulums of Milne and Von Rebeur-Paschwitz. The principal Italian instruments were of the vertical pendulum class. Their distinctive features are their capacity to register short as well as long oscillations, the resolution of the motion into its three co-ordinates, one vertical and two horizontal, and the use of a much more rapid motion of the paper ribbon receiving the traces, thereby giving full details of every vibration. The trace is made mechanically by a fine needle-point, or pin, instead of photographically by a light-ray falling on a strip of sensitised paper. It is, therefore, much sharper in definition. In the horizontal pendulum traces on slow-moving ribbons, the individual waves are crowded together, so as to be practically inseparable, while, in the vertical pendulum traces, the period, amplitude, and length of each wave are clearly shown.

A very large amount of study has been given to these seismograms by Professors Milne and Knott in England, by Drs. Agamennone and Cancani in Italy, by Professor Omori in Japan, and by Drs. Wiegand and Schlütter of Strassburg. Out of a vast mass of able discussion, we may

proceed to summarise the state of the subject to a comparatively recent date.

1. The tremors coming from a great and distant earthquake, recorded by delicate seismographs, were at first resolved into two groups, which were denominated preliminary tremors and large waves. The arrival of the preliminary tremors preceded that of the large waves by a time interval which increased with the distance. The following table may show approximately the time interval between the two arrivals at distances expressed in terms of arcs of a great circle.

INTERVALS BETWEEN ARRIVALS OF PRELIMINARY AND LONG WAVES

DEGREES OF ARC	TIME INTERVALS (Minutes)
20	5
40	11
60	20
80	29
100	38

2. The preliminary tremors were resolved by Dr. R. D. Oldham into two groups, and he showed that on many distant seismographs the two were more or less distinctly perceptible.[1] He suggested that the first of these groups to arrive consisted of normal waves which had passed through the earth by "brachystochronic" paths (*i. e.*, by paths requiring the shortest possible time, due regard being had to Schmidt's law of refraction caused by varying elasticity). The second group, he suggested, were transverse or dis-

[1] *Phil. Trans.*, vol. cxciv., 1900.

tortional waves. These, too, he thought, might have passed through the earth by brachystochronic paths. Dr. Oldham's views seem to have met with very general acceptance, and increasing observation brings more and more support to them. He thus points out three phases in the arrival of long-distance waves.

3. As shown by the Italian seismometrographs, the first phase consists of vibrations which are of very minute amplitude, having periods which seem to increase with the distance from the centrum. Thus, at 20° to 25° from the origin, the period may range between 1½ and 2½ seconds; at 40°, between three and six seconds; at 80°, between four and eight seconds. It is as yet uncertain how far this variation of period depends upon the initial energy and the impulse, and how far upon the different capacities of different instruments to pick up such very minute tremors. In almost every seismogram the beginning of the preliminary tremors is so gradual that there is more or less uncertainty whether tremors too faint to be caught by the instrument may not have preceded the earliest ones recorded.

4. The second phase is in many cases, perhaps in most cases, indicated rather distinctly. It follows the first by an interval which varies with the distance from the origin. The table on page 220 shows approximately the interval between the beginning of the first phase and the beginning of the second.

The amplitudes of waves of the second phase are notably larger than those of the first, and the wave periods are from fifty to eighty per cent. longer. The periods are also more regular.

INTERVAL BETWEEN BEGINNINGS OF FIRST TWO PHASES

Distance from Origin	Interval (Minutes)	Average Interval (Minutes)
20°	2 to 4	3.5
30°	3 to 7	5.
40°	4 to 8	6.6
50°	5 to 8	7.6
60°	6 to 9	8.5
70°	6 to 10	9.4
80°	7 to 10	10.
90°	7 to 10	10.5
100°	8 to 11	11.0

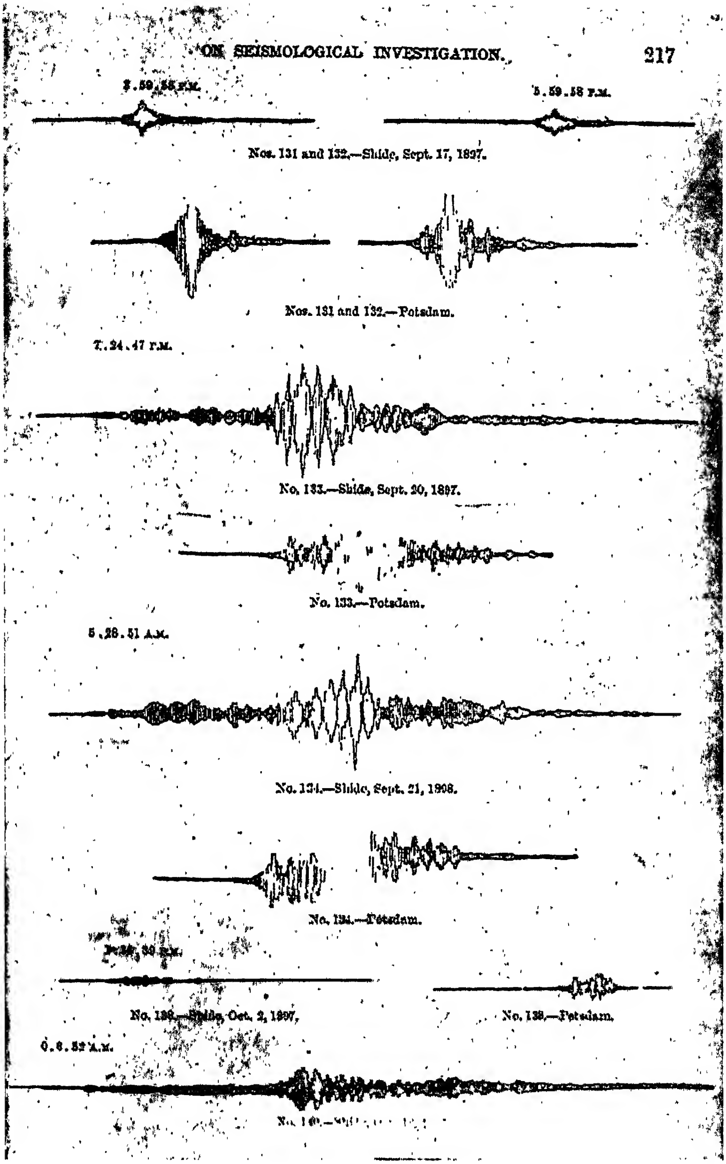
5. The third phase usually begins somewhat abruptly after the second phase shows signs of dying out. The amplitudes greatly increase, suddenly becoming four, six, or even ten times larger than those of the second phase. The periods also increase. From the two or three seconds of the first phase, and from the three or four seconds of the second phase, the period may become eight, ten, twenty-five, or even sixty seconds according to the distance from the origin. The periods also become remarkably regular. The accompanying seismograms from Italian observatories will indicate these phases.

The time relations of the several phases are shown graphically in Fig. 61, which is reproduced from Dr. Oldham's paper in the *Phil. Trans.*, 1900. With a much larger accumulation of observed data Professor Milne has recomputed the same curves and obtains a series of them, agreeing very well with those of Dr. Oldham.[1]

The most conspicuous feature of this diagram, Fig. 61, is

[1] *Brit. Assoc. Rep.*, 1902, p. 66.

PLATE VII.



Seismograms.

the straight line which represents the time-arc relation of the large waves of the third place. It means that these waves traverse equal arcs of a great circle in equal times, or that their speed of propagation is uniform, measured along

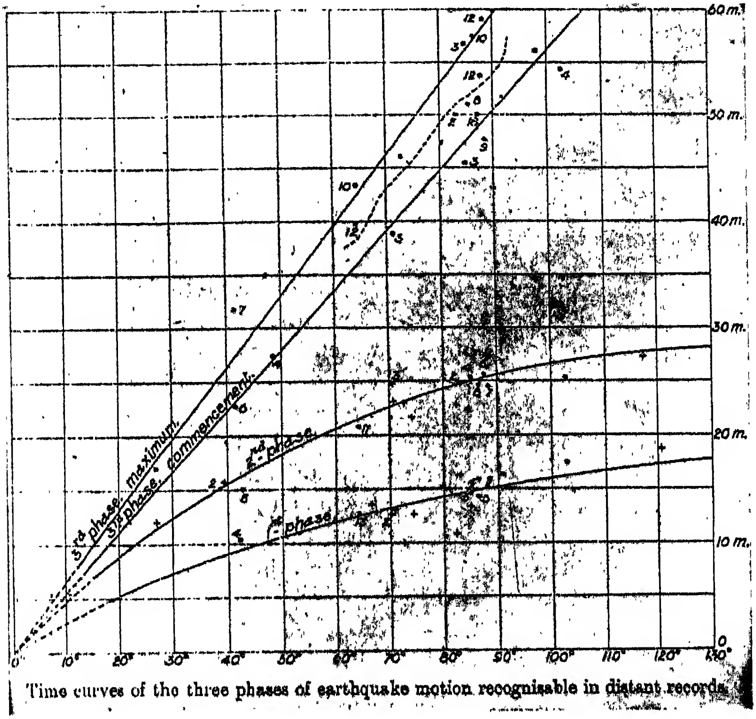


FIG. 61. Dr. Oldham's Time-Distance Relations.

the earth's surface. In terms of kilometres this is equivalent to about 2.95 kilometres per second.

The waves of the first and second phases show no such relation. As long ago as Professor Milne's first intimation of long-distance records the indications were that the

well-marked, long waves were preceded by tremors which travelled with much greater speed. As the observations of Von Rebeur-Paschwitz came to light and were followed by many others of the same nature, these preliminary tremors seemed to be more and more anomalous in respect to their speeds. As the seismographs accumulated it began to appear as if those tremors moved with a speed which bore a more definite ratio to the chord than to the arcs of great circles. But it was by no means a simple ratio. The relation was discussed tentatively by Professor Milne in 1897[1] and 1898. He says:

"If it is assumed that the motion (of the preliminary tremors) is propagated *round* the earth, then the velocities over arcs of 20°, 30°, 40° up to about 100°, which have lengths of 2200, 2300, 4000, and 11,000 kilometres, are about two, three, four, and eleven kilometres per second. Along wave paths less than 20° the velocity of two kilometres per second remains constant. For arcs greater than 100° the velocity apparently increases at a rate somewhat less than the rate at which the length of the arc increases. With the hypothesis that the vibrations travel along paths approximating to chords through the earth, then the above velocities must be reduced."

He then proceeds to tabulate apparent velocities of the preliminary tremors upon the hypothesis that they are proportional to the square root of the average depth of the chord. He finds an approximately simple ratio between the velocity and the square root of the average depth of the chord.

Prof. C. G. Knott, however, points out[2] that the relation

[1] *Brit. Assoc. Rep.*, 1897, pp. 173–174, and 1898, pp. 220–221.
[2] *Brit. Assoc. Rep.*, 1902, p. 67.

is more accurately expressed by the simple formula, that the time of transit is directly proportional to the length of the chord for the tremors of the first phase, but not so for the tremors of the second phase. Thus, in the following table, T is the time of transit to the antipodes and t the time recorded at the place of observation of the first phase, while T_1 and t_1 are corresponding times for the second phase. T is assumed to be 23 minutes, and T_1 33.4 minutes.

Arc	Sin ½ Arc	Ratio $\frac{t}{T}$	Ratio $\frac{t_1}{T_1}$	Remarks
20°	.174	.17	.20	It is to be understood that the chord is equal to the earth's radius multiplied by twice the sine of half the arc.
30°	.259	.26	.30	
40°	.342	.35	.39	
50°	.423	.44	.487	
60°	.500	.51	.57	
70°	.574	.58	.65	
80°	.643	.64	.72	
90°	.707	.70	.78	
100°	.766	.76	.83	
110°	.819	.82	.87	
120°	.866	.87	.91	
130°	.906	.91	.94	
140°	.940	.95	.96	
160°	.985	.98	.99	

Assuming that this table is fairly accurate its meaning is that:

1. *Tremors represented in the first phase traverse a diameter of the earth in about 23 minutes, while those of the second phase traverse the diameter in about 33½ minutes. The table assumes these times, but a glance at the diagram will show that the assumptions are fully justified.*

2. *The times occupied by waves of the first phase in traversing the several chords are in a simple ratio to the lengths of those chords. The speed of these waves, therefore,*

must be uniform along the chords, amounting to about 5.75 miles or 9.25 kilometres per second, to points between 20° and 180° of arc distant from the origin.

3. The times of transit by waves of the second phase are not simply proportional to the chords. Their speed appears to increase with the distance by some law not yet ascertained.

These results, especially the second, are remarkable and somewhat unexpected. Until 1891 long-distance observations seemed to be tending towards the inference that the speed of both normal and transverse waves increased with the distance from the origin. Such appeared a few years ago to be the impressions of Professor Milne, Drs. Agamennone and Cancani, of Rome, and of Professor Omori, of Tokio, all of whom had given the subject a very extended study. The above table of Prof. C. G. Knott, taken in connection with the graphic curves of Milne and Oldham, is, so far as I have yet discovered, the only summing up of the large mass of seismographic records which Professor Milne has with so much labour and ability gathered together during the last ten years. It appears that when these results are averaged up, the mean of all results for the waves of the first phase fits the theory of uniform speed well within the limits of probable errors of observation and fits it surprisingly well. Nor does it seem to fit any other.

If this result appearing in Professor Knott's table be accepted, it relieves seismology of some serious difficulties which have beset it sorely ever since Mallet's time. It leaves others untouched and brings up a new one which at first seemed to some investigators as formidable as any

which it solves. It relieves the subject of many of those perplexities arising from a variable speed of propagation which the elastic wave theory cannot explain without arbitrary assumptions, which are not only without support in observation, but which are intrinsically most improbable. The new difficulty which it introduces consists in the fact that the uniform speed thus deduced is nearly twice as great as theory had previously indicated from the measured values of the elasticity constants of the materials which constitute the great bulk of the earth-mass, as we know them.

An interesting result of the determination of the speed with which the surface waves are transmitted is the possibility of determining the place of origin of an earthquake which would otherwise be unknown because of its location beneath the seas or in the polar regions or elsewhere in some place so remote from observation that its impulses have dwindled into insensibility except to delicate instruments. When a great quake occurs in an inhabited region, even in Kashgar, Thibet, or Central Africa, its place of origin soon becomes known, and it often is possible to determine very nearly the moment of its occurrence. The speed of propagation to any distant observing station, then, becomes a mere question of time-observation. If, as appears now to be the truth, we find that this speed of transmission is practically uniform for the longer waves of all great quakes, we may, upon the assumption of uniform speed, trace them to their origin. It requires, however, at least four good time-observations at as many different stations. These stations also should be widely separated from each other.

Call these four stations A, B, C, D, in the order of priority in the time of arrival of the long waves. The speed of travel having been determined to be about 2.95 kilometres per second, the wave in one minute traverses about $1^{\circ}.6$ of the arc of a great circle. The difference in time of arrival between A and B in minutes multiplied by 1.6 will give in terms of degrees of arc the difference in their respective distances. With B as a centre draw upon a globe a circle whose radius is the same number of degrees. In the same way find from the time difference between A and C the corresponding radius, and with it describe a circle around C. Lastly, find the radius and describe the corresponding circle around D. Now find a circle which is tangent to each of these three. The centre of that circle is the place of origin.

CHAPTER XIII

DEDUCTIONS FROM SPEED OBSERVATIONS

The Physical Law Governing Wave Speeds — Investigations of Professor Nagaoka upon the Elasticity of Rocks—His Method of Determining Young's Modulus—Tables of his Results—Considerations upon the Elasticity of Deeply Buried Materials—Maxwell's Notion of Perfect Elasticity—Possible Approach to it at Considerable Depths—Hooke's Law and its Possible Continuity at Great Depths—The High Rigidity of the Earth's Interior—The Increase of Periods with Increasing Distance—Its Possible Explanation—Milne's Idea of Earthquake Echoes

BEFORE we can draw any inferences from the very important results set forth in the preceding chapters we must revert to the law which governs the speed of transmission. In doing so it will be necessary to assume that the reader has some elementary knowledge of the general theory and configurations of the principal kinds of wave-motion, and, therefore, of the subject of elasticity. Those who have not pursued this line of study may find some trouble in following the discussions.

It is understood at the outset that the speed of transmission of elastic waves is proportional to the square root of the elasticity and inversely proportional to the square root of the density of the medium; or, more briefly, is proportional to the square root of the elasticity-density ratio. In order to reach a numerical estimate it is necessary to know

the elasticity and density by measurement of samples, if they are attainable. Otherwise we are left to infer them by roundabout methods. The measurement of the density of a sample is the simplest possible procedure. The measurement of elasticity is far more complex.

Of the elasticity of the materials of the earth-mass we have very little knowledge. We could not, of course, have any direct knowledge of the elasticity of the deeply buried materials. But even the surface rocks have been seldom subject to the searching inquiries of the physicist so far as their elasticities are concerned, and the sources from which such information can be supplied are surprisingly scanty. There have been some investigations made with great care by Wertheim, Boussinesq, Cerruti, Lord Kelvin, and a few others in Europe, but the most extended ones have been made in Japan by H. Nagaoka, Professor of Applied Mathematics in the University of Tokio.[1]

His method consisted in determining the values of "Young's Modulus" and the "Rigidity Modulus" in selected specimens of rock from many geological horizons from the Archæan to the most recent. From the two the Compression or Volume Modulus follows at once by computation. The Young's moduli were obtained by preparing the specimens in bars fifteen cm. long and one cm. square, and subjecting them to flexure. The rigidity moduli were obtained by subjecting similar bars to torsion and deducing the rigidity by St. Venant's formula for the torsion of rectangular prisms.

[1] Publications of the Earthquake Investigation Committee in Foreign Languages, No. 4, Tokio, 1900.

Throughout the long series of these experiments the results in every case showed that Hooke's law did not hold good even for very small flexures and torsions. The deviation from strict proportionality between stress and strain was incomparably greater than in common metals. The after-effects when the loading or twisting was at all forcible were very considerable, and the return to the original shape was very imperfect. These low elastic limits made it necessary to confine the observations within correspondingly narrow limits of stress. In many rocks this imperfect resilience rendered it very doubtful whether any proportionality between stress and strain could be established.

A very marked difference was always found between the results of the same kind and degree of stress when applied to specimens cut parallel to the bedding planes and perpendicular to them. In each case specimens were cut in both ways and their moduli separately determined.

These results have been tabulated by Professor Nagaoka,[1] and they are so valuable that it is thought that they should be published here, as they may be found useful in many other relations. He incorporated in the tables computations of the speed of both normal and transverse waves for a medium composed of each kind of rock. The speed of the normal wave is computed by taking Young's modulus as the elasticity constant. The speed thus deduced is a little less than would be deduced for a wave in a medium having an unlimited extent in every direction, and is more

[1] Professor Nagaoka's paper and discussion is reprinted in full in the L. E. and D. *Phil. Mag.*, series v., vol. l., 1900, a series not found in all libraries by any means.

nearly that of a normal wave moving from end to end of a long bar. Thus the speed of a plane wave in an unlimited medium of steel would in theory be about 6.2 km. per second, while in a rather slender steel bar it would be about 5.3 km. per second, the ratio under the two conditions being 1 : .855. In the table every entry is reduced to the centimetre, gramme, second scale; Z = density, E_1 = Young's modulus perpendicular to the bedding planes, E_2 = Young's modulus parallel thereto, E = the arithmetical mean of the two, μ = rigidity modulus, V_1 = speed of normal wave, V_2 = speed of transverse wave.

NAGAOKA'S ELASTIC CONSTANTS OF ROCKS AND WAVE SPEEDS IN KILOMETRES PER SECOND

ROCK	Z	E_1	E_2	E	μ	V_1	V_2
Archæan:							
Chlorite schist........	2.977	112.1	132.4	122.3	24.03	6.40	2.84
" "	2.955	146.0	147.6	146.8	31.57	7.05	3.27
Eruptive:							
Peridotite / Serpentine	2.825	72.92	58.99	65.96	22.24	4.83	2.81
"	2.777	62.42	55.86	59.14	20.09	4.61	2.69
Ophicalcite...........	2.786	54.15	53.90	54.03	19.73	4.41	2.66
Peridotite............	2.593	38.90	53.71	46.31	......	4.22	.....
Serpentine...........	2.570	39.03	46.00	32.52	16.00	4 07	2.49
Palæozoic:							
Schalstein............	2.653	120.50	92.25	106.40	18.90	6.32	2.67
Clay slate............	2.149	79.69	83.29	81.49	28.06	6.16	3.61
Schalstein	2.768	70.02	95.00	82.51	25.36	5.45	3 03
"	2.772	97.90	103.30	100.60	21.25	6.02	2.77
Sandy slate...........	2.640	81.79	92.40	82.10	17.05	5.75	2.54
Clay slate............	2.674	98.00	83.09	90.55	13.79	5.82	2.27
" "	2.690	90.64	86.71	88.68	20.75	5.74	2.78
" "	2.708	51.92	62.26	57.09	20.74	4.52	2.77
Limestone............	2.630	84.95	88.45	86.20	29.83	5.74	3.38
"	2.653	80.20	86.61	83.40	31.00	5.60	3.42
"	2.682	68.86	79.55	74.20	21.71	5.26	2.84
Weathered / Clay slate	2.314	62.15	61.35	61.75	10.03	5.18	2.08
	2 304	56.83	58.90	57.87	8.85	5.01	1.96
Marble..............	2.654	76.00	63.72	69.86	30.11	5.13	3.37
"	2.625	63.53	46.20	54.86	28.60	4.54	3.45
Schalstein............	2.824	74.60	70.52	72.56	18.96	5.07	2.58
"	2.762	57.68	37.70	47.69	8.98	4.63	1.80

Rock	Z	E_1	E_2	E	μ	V_1	V_2
Palæozoic:							
Weathered }	2.316	39.44	35.27	37.36	4.99	4.02	1.47
Clay slate }	2.306	35.37	36.69	36.03	5.27	3.96	1.51
Marble	2.650	37.26	37.64	37.45	15.08	3.76	2.39
"	2.650	37.33	28.33	32.82	18.80	3.93	2.66
Clay slate	2.384	34.48	30.76	32.62	8.00	3.70	1.83
" "	2.392	30.64	30.35	30.50	8.54	3.57	1.87
" "	2.462	30.35	28.10	29.23	.....	3.45	1.71
" "	2.416	31.00	31.86	31.43	......	3.61	.....
Weathered }	2.503	12.45	12.20	12.33	4.60	2.32	1.36
Clay slate }	2.500	13.00	13.64	13.32	4.31	2 31	1.31
" "	2.490	12.72	12.26	12.49	6.59	2.24	1.63
" "	2.500	12.54	12.47	12.51	4.43	2.24	1.33
Palæozoic, Eruptive:							
Granite	2.572	37.91	46.71	42.31	18.43	4.05	2.68
"	2.550	31.42	....	....	13.99	3.51	2.34
"	2.549	18.83	20.43	19.63	6.89	2.78	1.64
"	2 500	14.84	15.12	14.98	5.05	2.42	1.40
"	2.503	15.23	9.73	12.48	5 47	2.22	1.48
"	2.530	11.97	9.89	10.93	4.43	2.08	1.32
Mesozoic Rocks:							
Sandstone	2.216	9.2	9.0	9.12	3.1	2.03	1.18
"	2.236	7.1	7.2	7.12	2.4	1.78	1.04
"	2.223	7.7	7.6	7.67	2.7	1.86	1.10
Schalstein	2.778	75.7	83.0	79.4	23.2	5.35	2.89
Clay slate	2.711	88.4	99.3	98.8	22.6	5.88	2.89
" "	2.702	83.6	85.3	84.5	18.5	5.59	3.17
" "	2.681	32.2	50.6	41.4	14.8	3.91	2.35
" "	2.678	43.7	44.3	44.0	14.2	4.06	2.31
Cenozoic Rocks:							
Rhyolite	2.316	32.1	17.5	24.8	14.00	3.24	2.46
" Tuff	2.346	21.9	21.5	21.73	9.32	3.05	1.99
" "	2.316	21.8	20.0	20.90	8.05	3.01	1.86
" "	2.305	20.6	21.1	20.8	8.74	3.02	1.95
" "	2.321	21.2	21.4	21.3	8.45	3.02	1.91
Rhyolite	2.472	21.3	18.7	20.0	8.57	2.85	1.86
"	2.454	19.5	18.3	18.9	9.15	2.78	1.93
" Tuff	2.228	18.8	19.9	19.3	6.9	3.00	1.79
" "	2.198	17.4	11.8	14.6	......	2.59	.....
Rhyolite	1.945	11.3	11.7	11.5	5.78	2.43	1.72
"	1.944	14.0	15.1	14.6	5.86	2.74	1.74
" Tuff	1.889	8.11	10.1	9.1	4.2	2.20	1.49
Sandstone	2.345	10.9	11.4	11.2	4.6	2.18	1.40
Rhyolite Tuff	2.263	8.0	7.59	7.8	3.59	1.86	1.26
" "	2.228	10.8	11.1	10.96	6.25	2.22	1.51
" "	2.198	9.8	9.6	9.67	5.66	2.10	1.67
" "	1.371	1.43	2.49	1.96	1.06	1.19	0.89

All the figures in the four columns E_1, E_2, E, and μ must be multiplied by 10^{10} if the speed is desired in centimetres per second, or by 10^{6} if the result is desired in metres per second.

A marked feature of this table is the gradation in the values of the elasticity constants from the Archæan to the Quatenary. The older rocks in the table have, as a rule, much higher elasticities and slightly higher densities than the younger ones. The explanation suggests itself at once. When these older rocks were first deposited they probably did not differ from corresponding deposits of later ages. The difference has been one of the results of metamorphic or metasomatic action in the course of time, aided by the pressure of the masses which once rested upon them and which have been denuded. The probable meaning of this is that none of the rocks visible at the surface are in a condition to manifest the elasticity and rigidity they would have under such conditions as exist in the depths of the earth. The specimens which reach the hand of the experimenter are far from isotropic (*i. e.*, homogeneous throughout). They are usually aggregates of small crystals with their cleavage planes and cracks held in a glassy base, sometimes holocrystalline, but abounding in minute vesicles, microscopic cracks, and flaws, and with some crystals harder or softer, stronger or weaker, more or less elastic than the mean. The result of all this is to reduce the mean value of the elasticity in comparison with what would appear in a perfectly clear, compact mineral substance with a feebly pronounced cleavage like a quartz, topaz, or feldspar crystal. Towards this more compact and perfectly continuous condition the pressure of great depths in the earth should, it may seem, tend to bring the materials which are subject to it.

To the actual granular, crystalline condition with its

accompanying discontinuities and want of isotropy we may readily attribute the disagreement with Hooke's law which Professor Nagaoka found.

The view here suggested is that the deeper the materials are placed in the earth the nearer is the approach to "perfect elasticity," and for that term we may adopt Maxwell's definition, "A perfectly elastic body," says Maxwell, is one "which, subjected to a given stress at a given temperature, experiences a strain of definite amount, which does not increase when the stress is prolonged, and which disappears completely when the stress is removed." A perfectly elastic body is also defined by Thompson and Tait as one which "when brought to any one state of strain requires at all times the same stress to hold it in this state however long it may be kept strained, or however rapidly its state may be altered from any other state of strain, or from no strain, to the strain in question."

Hooke's law, *ut tensio sic vis*, involves either or both of these definitions as a major premiss, and wherever it fails it does so presumably because the body is not continuous in all its parts, or because at points within its mass the body has been strained beyond the elastic limit. At a depth far below human access, but very small relatively to the length of the terrestrial radius, these defects of isotropy and continuity may be so far compensated as to bring the material to the highest degree of elasticity of which it is capable. Along the line of descent towards this depth the elasticity, according to the foregoing view, should increase more rapidly than the density. Below that depth Hooke's law may hold good.

This suggestion seems to be the interpretation of Professor Nagaoka's tabulated results so far as concerns the depths represented by the rocks he has tested. The densities increase but little towards the depth at which they acquired their present mineral and lithological conditions, but their elasticity and rigidity constants increase much. In fact, the Archæan schists at the beginning of the table have constants nearly high enough to satisfy the requirements of a wave speed of 9.25 km. per second. For if we compute from them the wave speeds according to Lamé's or Lord Kelvin's formula for a wave in an unlimited medium instead of the formula for a bar, as Professor Nagaoka has done, we find for the two schists speeds of 7.4 km. and 8.2 km. respectively. It would require an increase of only about twenty per cent. in the two elastic constants to account for a speed of 9.25 km. per second. This is far from incredible, and, indeed, presents no apparent difficulty.

The case of the tranverse waves forming the second phase in the distant seismograms is different. Assuming that these also travel along chords, the table of Professor Knott shows that their speed increases as they approach the centre of the earth and diminishes as they travel away from it. As these waves involve rigidity and density alone, the inference seems to be that the rigidity-density ratio increases a little toward the earth's centre. This is in no way surprising. Experiments on the flow of solids, though few in number, tend to the same conclusion, that under increasing pressure their rigidity increases faster than their density. It is one more contribution to the evidence that the effective rigidity of the earth's interior is very great.

Plate VIII.

Fig. 1.—Rocca di Papa. Pendolo orizzontale E.-W. Terremoto delle Indie, Provincia di Assam e Bengal, 12 Giugno, 1897.

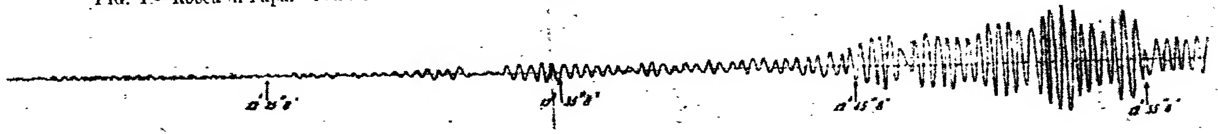


Fig. 2.—Terremoto delle Indie. Rocca di Papa, 12 Giugno, 1897. Pendolo orizzontale N.-S.

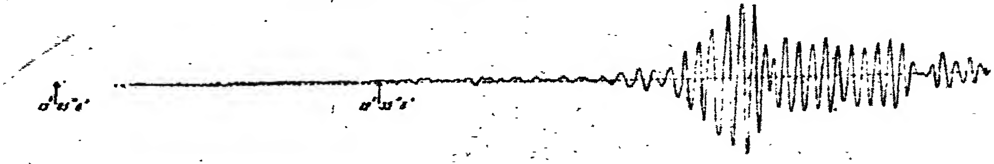
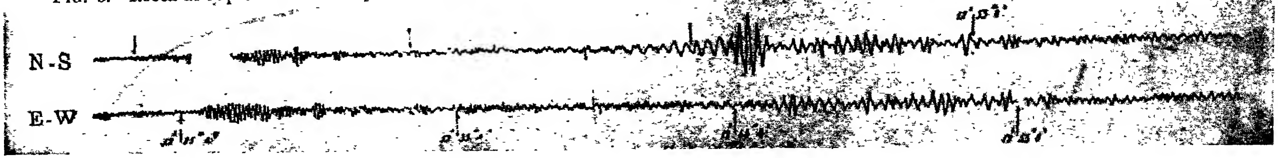


Fig. 3.—Rocca di Papa. Sismometrografo di 15m. Terremoto delle Indie, Prov. di Assam e Bengal. 12 Giugno, 1897. (Adolfo Cancani.)



Seismograms of the Great Assam Quake Taken at the Rocca di Papa Observatory near Rome.

There remains one phenomenon brought to light by the horizontal pendulums which presents a more serious difficulty. This is the lengthening of all the wave periods and also of the total duration of the disturbance as the distance from the origin increases. The theory of wave-motion in an unlimited isotropic medium furnishes no explanation of this phenomenon. The theory leads to constancy of period only, just as sound-waves in air are propagated without change of pitch (*i. e.*, period) for indefinite distances. Since we find no explanation consistent with an unlimited isotropic medium, we seem to have no resource but to look for it in the medium considered as limited by the terrestrial surface, and as heterogeneous from many causes. For the present, however, we have only to accept the fact and patiently await its explanation by the future results of observation and analysis.

It is possible that the suggestion of Professor Milne if followed up might ultimately lead to some light on this question.[1] He says:

"An earthquake disturbance at a station far removed from its origin shows that the main movement has two attendants,[2] one which precedes and the other which follows. The first of these by its characteristics indicates what is to follow, whilst the latter in a very much more pronounced manner will often repeat at definite intervals, but with decreasing intensity, the prominent features of what has passed. Inasmuch as these latter rhythmical but decreasing impulses of the dying earthquake are more likely to result from reflection than from interference, I have provisionally called them *Echoes*."

[1] *Brit. Assoc. Rep.*, 1899, p. 227.

[2] Professor Milne here considers as a single phase those two sets of preliminary tremors which have just been discussed as the first and second phases.

An examination of the different seismograms, Pls. VII and VIII will indicate that the third-phase pulsations of large amplitude reach a maximum, then rapidly decline, and some minutes later are repeated on a diminished scale of amplitude. There may be several repetitions, each succeeding one being less pronounced than its predecessor. Professor Milne continues:

"The main point at issue, and the one to be answered before we enter into further speculations, is whether seismograms showing this musical-like repetition can be interpreted in the manner here suggested. The concluding vibrations of an earthquake have usually been regarded as a disorderly mob of pulsatory movements resulting from spasmodic impulses which gradually grow feebler as the activity at a seismic centre became exhausted. The question before us is whether an earthquake dies by a process analogous to repeated and irregular settlements of disjointed materials, or whether it is a blow or blows which come to an end with musical reverberations inside the world. For the present my opinion inclines to the latter, and I see in the earthquake followers the likeness of their parent."

CHAPTER XIV

SEISMIC DISTRIBUTION AND CHARTING

The Catalogues of De Montessus de Ballore—His Conception of Seismicity—Frequency and Average Intensity—The Three Categories of Data: Historic, Seismologic, and Seismographic—Difficulties in Handling Frequency Data—Difficulties of Intensity Averages—De Montessus's Method of Avoiding them—His Numerical Measure of Seismicity—Subdivision of his Catalogue—Projecting Seismicity upon Maps—Mallet's Method—Locations by Epicentres—Dr. Davison's Method—His Geography of Japanese Quakes, and the Distribution of Milne's Catalogue—Tendency to Great Detail in Charting Seismic Distribution—Seismic Topography

THE widely varying degrees in which different portions of the earth are affected by seismic action have doubtless attracted general attention. All that is called for here is to give a sufficiently detailed statement of the extent to which seismic action varies and its geographic distribution.

This branch of the subject has received far more attention and study from M. De Montessus de Ballore than from any one else—much more, indeed, than from all other seismologists together. His work has been chiefly the compilation and treatment of very large masses of data consisting of catalogues of earthquakes in many parts of the world. They include also all authentic reports, from whatsoever source, which are sufficiently detailed to be of any use. His inquiry into the validity of Perrey's laws relating to the effect

of the moon's attraction upon seismic frequency has already been referred to. In the present chapter an account will be given of his investigations of the subject of "Seismic Geography," as he terms it, which is a name for the geographic distribution of seismic frequency and intensity. The words "frequency" and "intensity" obviously imply some kind of numerical measure, and to this measure he gives the name *seismicity*.

Every part of the earth's surface which is in any degree, however small, affected by earthquakes may be said to have its own seismicity, just as every region, unless absolutely rainless, has its degree of annual precipitation which is capable of a numerical statement in terms of cubic inches of rainwater falling annually upon each square inch of surface. We may speak of the seismicity of a region just as we speak of its vulcanicity, meaning in either case the degree to which it is affected by seismic or volcanic action. Scientific accuracy, however, demands a full understanding of just what quantity or quantities the term seismicity is intended to include and in what relations.

It is clear that it should include the frequency, *i. e.*, the number per day, week, or year, and the average intensity of the quakes or "*seisms*" as De Montessus terms them. It is equally plain that there must be a very definite understanding as to the area over which the frequency is to be distributed. With a full knowledge and definite understanding of these three quantities, frequency, intensity, and area, it becomes possible to establish a numerical measure of the seismicity of a district.

But it is not so easy a matter as it might at first seem to

determine satisfactorily either the frequency or the average intensity. It is obvious that the first requirement for determining frequency is a knowledge of the number of quakes occurring in a given period of time. For some districts, the number of which has been steadily increasing from decade to decade during the last thirty years, we have records of varying degrees of fulness, though it is probable that not more than a dozen localities have complete records including all of the quakes. For the greater part of the world the record is more or less fragmentary, and for a very considerable portion there is practically none at all, or next to none.

De Montessus divides the sources of information into three classes: (1) Historic, embracing publications of all kinds other than strictly seismologic. These are histories, narratives, newspapers, etc., whose common feature in this relation is that their accounts are incidental only and not for scientific or statistical purposes. It is quite certain that they mention only a small fraction of the entire number of quakes in any region. (2) Seismologic observations, which are made for the purpose of recording all the *sensible* quakes of a region, *i. e.*, quakes which are perceptible to the human senses without the help of instruments. They are made by individuals, or by associations, for the purpose of promoting knowledge of seismology and for making it as full and accurate as practicable. (3) Seismographic observations, which are made in observatories, public or private, equipped with delicate instruments, which record all vibrations, great and small.

It might seem as if the seismographic class of records should be much more satisfactory than the others whenever

they are obtainable. De Montessus thinks otherwise. Seismographs record not only true earthquakes but pseudo-quakes also, *i. e.*, tremors produced by agencies not of a seismic nature at all, such as the passage of railway trains or carriages or the effect of wind on neighbouring buildings or even upon hills, the beating of the surf, etc., and it is sometimes a difficult matter to distinguish such records from those of true earthquakes of low intensity. The seismograph also records the dying tremors of quakes originating far outside the district under consideration, and it is equally difficult to separate them from tremors originating within the district. Seismographs are usually installed within, or near, large towns or cities, and for various reasons fail sometimes to pick up the faint tremors caused by small quakes far out in the country, though still within the district, thus giving undue frequency to urban, at the expense of rural, localities. For these and other reasons he regards seismographic records as less desirable than the seismologic as sources for the data upon which frequency estimates should rest. Their errors are always in excess. They give records which should be rejected, but which are difficult to distinguish from those which should be retained.

Seismologic records, he fully recognises, are always defective in the opposite direction, as they fail to note a large number of small quakes. This deficiency, however, he thinks, can sometimes be supplied, and his method of doing so is certainly an ingenious one. In his vast catalogue of earthquakes he finds a considerable number of districts, ninety-four in all, which have at least two of the three kinds of records. These enable him to establish ratios between

the frequencies resulting from the three kinds of data. Thus if h be the mean annual number of quakes obtained from historic sources, l the mean annual number from seismologic, and g the same from seismographic records, the following ratios are given as the results of comparisons in the ninety-four districts:

From	44	districts	$l/h = 4.26$
"	28	"	$g/h = 26.59$
"	22	"	$g/l = 6.44$

These values are each determined empirically and quite independently of each other. If either of them is computed by substitution from the other, the agreements will be found to be remarkably close. This gives him confidence in their accuracy. These are ratios, however, and not absolute values. To ascertain those absolute values and determine what proportion of quakes escapes detection it is necessary either to assume that the seismographic observations are practically complete, *i. e.*, include all the quakes, great and small, or else find some independent means of determining the absolute values of one of the frequencies. If it be granted that the seismographic records of a district are complete it follows from the above ratios that the historic and seismologic records show respectively only 3.76 and 15.52 per cent. of the shocks and lose 96.24 and 84.48 per cent. of them. Presumably those which are lost are in much the greatest part quakes of small intensities, and those which are secured are of the higher intensities.

It is to be feared that this method of computing seismologic frequency from either historic or seismographic data is much too hazardous. The fulness with which earthquakes

are recorded historically and seismologically must vary extremely as between different countries, and any attempt to amend a defective record by the application of a systematic rule seems to be of doubtful utility. The most that can be said is that in districts where there are no seismographic and only very imperfect historic or seismologic records the method might be employed for temporary estimates only, but with great caution, pending the acquisition of more authentic data.

Intensity estimates also have their difficulties. For distributive purposes it might at first sight seem as if each quake should have its own intensity estimate for which the Rossi-Forel scale, or its equivalent in some other conventional scale, must be used, except in those rare cases where seismograms give the means of making an absolute determination. But De Montessus maintains that this is not necessary. In an interesting paper in *Bolletino Soc. Sis. Ital.*, vol. iii., he discusses Milne's catalogue of Japanese quakes, 1885–1892. He first points out that in powerful quakes the best measure of their intensity (*i. e.*, total energy) is the area which is sensibly shaken. He then shows that the intensity of a single small quake cannot be estimated in that way, being dependent upon the depth of its origin. But in a region where they are frequent, and the depths variable, the average intensity can still be measured by the average area of sensible vibration. He finally shows that in such conditions the mean intensity varies in the same way as the frequency: in other words, when it quakes often the shocks extend over a greater mean area, *i. e.*, they are on the average more intense, and *vice versa*.

"Since, then, the frequency and intensity vary, broadly speaking, in the same way, we have the very interesting result that, to express the relative importance of the quakes of a district, *i. e.*, its seismicity, we only need to consider their mean frequency." This statement may be regarded as broadly correct, and it greatly facilitates the way toward a numerical measure of seismicity.

The extent of country involved in any group of quakes also requires definition for each district; for seismicity, considered as a numerical measure, means the frequency and intensity of seismic action per unit area. It is clear that it makes a very important difference whether this area is to extend over an entire state, or is to include only those portions of it within which the sensible quakes occur in compact groups; whether it is to include the whole marine and insular areas of an archipelago or is to cover only the islands. The question here raised is of importance only when it becomes desirable to segregate some particular district. The limits in such a case must be selected more or less arbitrarily, and De Montessus thinks it best to include the marine areas of archipelagos and also to include such portions of the sea adjoining peninsulas or shore lines as are proven to have had actual epicentres. Such is also the prevailing opinion among seismologists.

We may now pass to his method of establishing the numerical measure of seismicity. If, in a region whose surface is A in square kilometres, there be observed n quakes in p years, the mean annual frequency for the region will be $f = n/p$, and f/A will be the mean annual frequency per square kilometre. The reciprocal of this last quantity,

A/f, is the fractional area into which the entire area A may be divided up so as to give to the fractional area one quake per annum. As this is a surface quantity, and as a linear quantity is on many accounts preferable, we may consider this fractional area as a square and take its square root, $\sqrt{A/f}$. This is De Montessus's seismicity measure or measure of the relative importance, *i. e.*, frequency and intensity, of earthquakes in a given district. Translating algebra into prose, its meaning is that a given area affected by many earthquakes may be conceived of as being divided into parts of such size that if all the earthquakes were divided up equally among them there would be exactly one quake per annum for each part. The parts being small equal squares, the side of the square is taken as the measure. Or again, the region is divided up into as many small equal squares as the average number of quakes per annum. The side of one of those squares is the measure of the seismicity.

It may be a somewhat awkward one at first and until it becomes familiar by use. For it is a reciprocal yard-stick, *i. e.*, the smaller the measure the greater is the quantity measured, and *vice versa*. But it is scientifically sound and rational and as definite as the facts admit. It considers averages or arithmetical means. The number of quakes varies from year to year, and there is no resource but to take the mean of a series of years, the longer the better. For though two consecutive years may differ considerably, the mean of a dozen years or more is much more reliable. There is, however, an important class of exceptions. A great tectonic quake of the first order of magnitude is invariably followed by a long series of after-shocks diminish-

ing in time but recognisable sometimes for ten or fifteen years after the main convulsion as its logical consequences.[1]

The work of De Montessus has for its principal object the determination and charting of the distribution of earthquakes. He has undertaken a task of impressive magnitude, and his results have thus far been of great value and importance to seismology. The first thing needed was a great catalogue of earthquakes scientifically observed, and in the course of twenty years he has accumulated one embracing 131,292 quakes and 10,499 epicentres to January 1, 1900. This is a body of observed facts in comparison with which the catalogues of Perrey, Mallet, Fuchs, and Von Hoff look small indeed. But, what is of much more importance, they include a far greater proportion than any of the older ones of quakes which have been scientifically observed according to modern standards, and which give information in detail of the kinds that are most needed for comparisons and estimates, and in which the records of the older catalogues are very deficient.

A preliminary synopsis of his greater work in this field has been published.[2] It divides the seven grand divisions of the world into fifty subdivisions or "chapters" and 451 seismic "regions."[3] These are arranged in tabular form, showing in each region (district) the number of epicentres,

[1] It is a doubtful question whether the after-shocks of the New Madrid earthquakes of 1811–12 have entirely ceased even yet. At all events, this locality is certainly a seismically sensitive one to-day and has been so in a diminishing degree ever since 1812.

[2] In *Beiträge zur Geophysik*, vol. iv. "Introduction à un essai de description sismique du globe et mesure de la sismicité."

[3] Where he uses the term "region" I have used the term "district," which is believed to be a better translation of De Montessus's real meaning.

the number of quakes, the period within which they were observed, and, if the data are sufficient, the frequency and the computed seismicity. He omits to state the areas adopted in the computations and these the reader is left to compute for himself. This table is followed by a second in which the 451 regions are entered in the order of their seismicity. As it may be of interest to see how the seismicity of the different parts of the world works out, these tables are given in an appendix. It is necessary to recall that, the measure being a reciprocal one, the smaller it is the higher the frequency and intensity of the earthquakes.

To Dr. Robert Mallet is due the credit of having begun the transformation of the contemplative and speculative stages of seismology into the scientific stage. In the attempt to systematise existing knowledge one of his efforts was to produce a graphic representation of the distribution of earthquakes over the globe. In the *British Association Report* for 1858 he published a small-scale copy of a map he had prepared with that view. His general plan was to colour the area affected by each quake with a tint whose darkness should be proportional to the intensity. He divided earthquakes according to their energy into three groups, the first consisting of those of greatest power, the second of mean power, and the third of minor quakes. The depths of the tints were proportional respectively to the numbers 9, 3, and 1. If the data were insufficient to fix definitely the areas shaken, their radii were assumed in the above proportions of 9, 3, and 1. As a repetition of a quake in any region increased the tint the darkness of the colour might represent either frequency or intensity or both.

Although Mallet's graphic method was adopted by many subsequent investigators it lacked two features which recent seismological requirements demand. It failed to show the points at which the disturbances originated. At the present time it is considered necessary that seismic maps should show, if possible, the locations of epicentres. It is true that there are very few countries in the world where this is now possible. But the growth of interest in the subject is bringing us nearer every year to the conditions in which it will be possible to extend this method to a greater and greater number of countries. Another requirement of the new seismology in the matter of graphic distribution is a much fuller detail than was possible in Mallet's time. The projection of seismic areas of great extent upon a small-scale map is apt to prove deceptive and to encourage broad generalisations which a more detailed and accurate representation is almost certain to dispel. Mallet's map is a case in point.

His distribution shows dark bands surrounding the greater marine basins, and appears to have a close association with the volcanic areas. The areas of least seismic activity, he indicates, are "the great oceanic or ter-oceanic basins, and the large islands existing in shallow seas." The association of seismic and volcanic distribution has already been adverted to (Chapter III). With regard to the seismicity of oceanic basins and of large islands in shallow seas it has been established by long observation since Mallet's time that earthquakes are as frequent and as powerful, if not more so, in those basins as in land areas, and that islands in shallow seas are no more exempt than continental regions.

In truth, with recent accumulations of observed data the favourite breeding-grounds of tectonic earthquakes are shown by De Montessus to be the regions where the variations of profile are greatest. The distribution of volcanic quakes will have intelligible meaning when we arrive at some coherent, rational theory of volcanic causation and not before. At present we have none which is at all satisfactory.

By far the best, perhaps it might be said, the only large-scale map showing the distribution of epicentres is Milne's seismic map of Japan founded upon the observation of 8330 quakes in the period from 1885 to 1892. This great catalogue, probably unequalled in the thoroughness with which it was compiled, gives, with very few exceptions, the epoch, the dimensions of the area of sensible disturbance and its boundary, and the position of the epicentre of each disturbance.

In projecting this catalogue upon the map the empire is divided by meridians and parallels into more than two thousand rectangles having sides ten minutes in both length and breadth and consecutively numbered. In the reduced copy of the map[1] the positions of the epicentres are shown by small dots, though in some of the districts they were so numerous that there was insufficient room for all, and the boundaries are represented as straight lines beyond their proper limits.

Dr. Charles Davison has improved much upon this reduced chart of Professor Milne, which does not show clearly

[1] Published in the *Seismological Journal of Japan*, 1895, and reproduced in *Nature*, 1897.

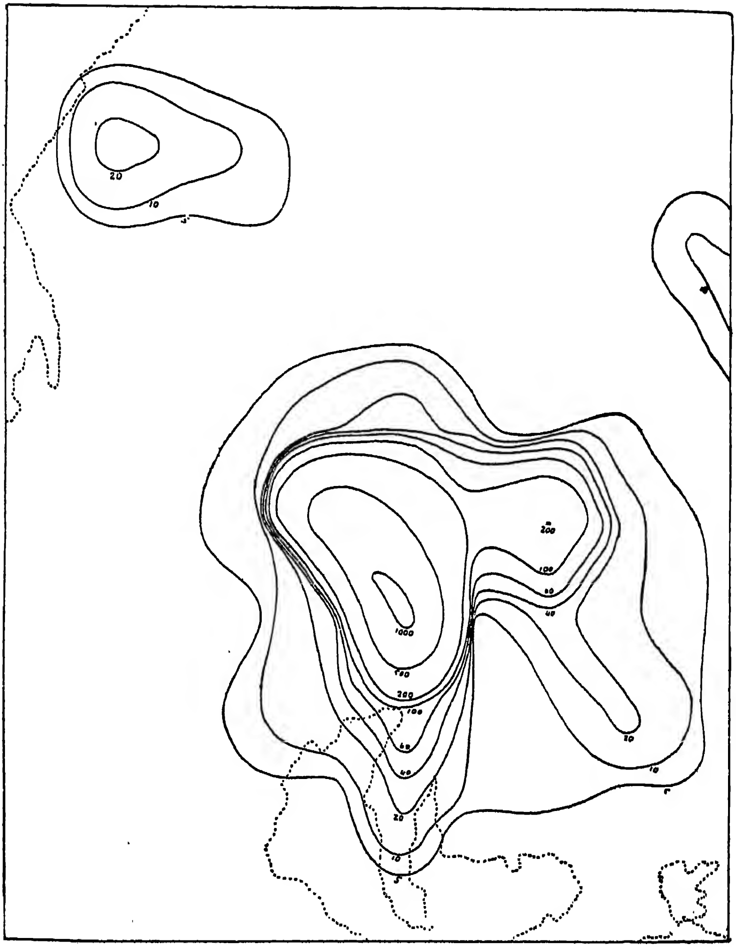


FIG. 62. Dr. Davison's Method of Representing Frequency.

the full details of the distribution and does not indicate the gradation of seismicity. This lack Dr. Davison has supplied in a very gratifying way.[1] His method consists in drawing curves through the centres of all rectangles in which the same number of epicentres occur. These curves are therefore analogous to topographic contour lines or to isobasic or isothermal lines. When they are close together they imply a rapid change or gradient in seismic activity; when widely separated they imply a slow gradient. Fig. 62 is a limited district in the provinces of Mino and Owari bounded by the parallels 34° 40′ and 36° 20′, and by the meridians 2° 10′ and 3° 50′ west of Tokio. The curves correspond to 5, 10, 20, 40, 60, 100, 200, 500, and 1000 epicentres in a rectangle. Pl. IX is a seismic map of the entire country. The curves correspond to 1, 5, 10, 50, 100, and 500 epicentres in a rectangle. The intervals between contours are also shaded to show the distribution more distinctly.

Dr. Davison remarks that the interest and value of the map would have been much increased if the curves had been continued over the surrounding sea. But it seemed undesirable to attempt this for several reasons. It is difficult, for instance, to determine with accuracy the position of the epicentre when only part of the boundary of the disturbed area can be drawn. Another and more important reason is that, while some earthquakes are known to have originated at a distance as great as fifty or sixty geographical miles from the coast, it is only the stronger of such earthquakes that can be felt on the land at all, and still

[1] *Journal of the Royal Geographical Society*, Nov., 1897. I avail myself freely of this very instructive article and its accompanying maps.

fewer that can be felt over an area large enough for the epicentre to be located even approximately. Thus the frequency curves would have been accurate only in the immediate neighbourhood of the coast.

Dr. Davison's map, Pl. IX, presents to the eye in a very striking manner the truth of Professor Milne's remark that his "map of earthquake origins or centres shows that the central portions of Japan, which are the mountainous districts where active volcanos are numerous, are singularly free from earthquakes." The volcanos are indicated by dots and seldom occur in the darkly shaded areas.

Although the most recent tendency in treating of seismic distribution is toward the development of details as illustrated by Dr. Davison's and Major De Montessus de Ballore's work in the preceding chapter, there is still a reluctance to abandon altogether the older method of grouping seismicity in large regions, each covering considerable fractions of the earth's surface. It is the method of Mallet, Perrey, and their immediate successors. It is by no means without value. Milne, though one of the chief advocates of the newer detailed method, has recently resorted to it, and on a Mercatorial map of the world has outlined no less than twelve great seismic regions,[1] each covering a million square miles, or more.

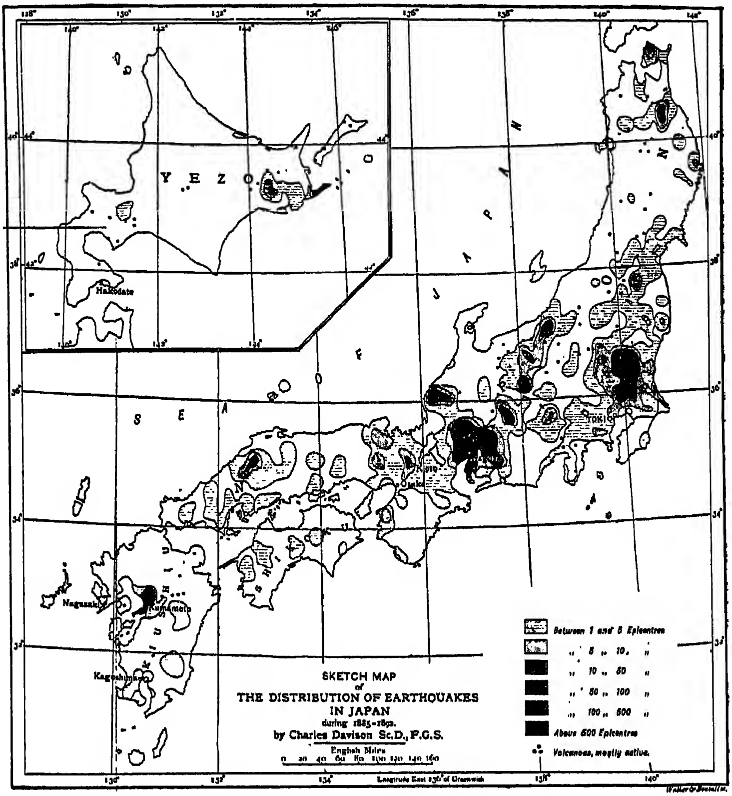
In delimiting regions of such vast extent characterised by great seismic frequency, the natural tendency of the modern seismologist is to inquire into the details of seismic distribution and to look for subdivisions. He cannot suppose that this distribution is uniform, and may ask if this great

[1] Published in *Brit. Assoc. Rep.*, 1902.

expanse is really one immense field possessing a well-defined unity or is composed of several fields which are quite distinct in origin or causation, and having no assignable relation to each other except their propinquity. But propinquity, though it may suggest the possibility of interrelation, does not necessarily mean anything; and to know whether it has any meaning we are led at once to look more closely into details. And generally when we do so we find that these vast regions are really composed of distinct fields or districts having no apparent interdependence.

All of these great seismic regions are with one exception areas in which the sea predominates over the land. The single exception is the long belt of mountainous country bordering on the north of the Mediterranean, including the Italian and Balkan peninsulas, Asia Minor, and extending thence eastward through Persia and Turkestan to the Pamirs and Himalayas. The enormous extent of this belt ought to suggest at once that it cannot be a single seismic region. The most cursory examination must lead to its subdivision into parts. Southern Italy has no assignable seismic relation to the Piedmont, the Riviera, the Swiss Alps, or the Austrian Tyrol. Nor have these subdivisions any such relation to the seismic districts of the Balkans, the Ionian Islands, Thessaly, Greece, the Ægean Archipelago, or the western coast of Asia Minor. A wide interval in which quakes are not common separates the seismic districts bordering on the Propontis and Ægean from those of the Lebanon and Taurus ranges and from the severely shaken valleys and plains of Armenia and the sources of the Euphrates and Tigris. The seismic districts of Turkestan are

PLATE IX.



YEZO
Hakodate
SEA OF JAPAN
Nagasaki
Kumamoto
Kagoshima
TOKIO
SKETCH MAP
of
THE DISTRIBUTION OF EARTHQUAKES
IN JAPAN
during 1885-1892.
by Charles Davison Sc.D., F.G.S.
English Miles
0 20 40 60 80 100 120 140 160
Between 1 and 5 Epicentres
„ 5 „ 10 „
„ 10 „ 50 „
„ 50 „ 100 „
„ 100 „ 500 „
Above 500 Epicentres
Volcanoes, mostly active.
Longitude East 130° of Greenwich
Walker & Boutall sc.

separated by wide intervening regions of comparative quiet from all other disturbed areas.

If we were to consider each or any one of the districts suggested in the preceding paragraph in detail, and if we had as full a knowledge of the facts as we possess of Japan or Italy, and if we should project upon the map of the district its seismicity in the same manner as Dr. Davison has done Japan, we should have a map resembling in general character the one he has drawn. The two would have the same general resemblance as two turns of a kaleidoscope. There would be localities in which shocks are very frequent and epicentres densely crowded together. The contour lines would be very near each other in some places and wide apart in others. The significance of the shading of the map would still be a problem. The reason why it should be darkly shaded in some spots and lightly shaded in others would still remain to be sought.

CHAPTER XV

SEISMIC REGIONS AND GEOGRAPHY

Value of Studies in Distribution Derived from Relations to Other Distributive Facts—Relations to Vulcanicity—Independence of Seismic and Volcanic Centres in Japan—Also in the Philippines—In the Kuriles and Aleutians—On the Alaska and California Coasts—Tectonic Nature of the Quakes in those Regions—Southern and Central Mexico—Volcanic Nature of their Quakes — Central American Volcanos and their Several Subdivisions—Relation of Seismic Activity—Close Proximity of Epicentres to the Volcanos and their Interdependence—Shallow Foci of Central American Quakes — Peruvian Quakes — Chiefly Tectonic—Chilian Tectonic Quakes—Frequent Submarine Origin of Them—New Zealand Quakes—Their Investigation—Low Degree of Seismicity in Australasia as a Whole—De Montessus's Views of the Relations between Seismicity and Topographic Relief—Stable and Unstable Regions—Instability of Great Topographic Slopes—De Montessus's Statement of the Laws of Relief and Seismicity—Illustrated in the West Indies and in the East Indies

THE study of the geographic distribution of a category of facts must derive its chief value from the relations to other facts which are thereby disclosed, and especially from their relation to possible causes. We have already considered earthquakes as divisible into two classes, volcanic and tectonic, according to their causes. We cannot, it is true, always declare with confidence to which of these classes a particular quake should be assigned. But in most cases we can show a preponderance of probability in favour

of one or the other. One of the significant facts whenever the question of class arises is the situation of the epicentre with respect to an active or recently extinct volcano. Close proximity to a volcano raises a presumption that the quake is of volcanic origin. Tectonic quakes also have, as has been already pointed out, certain characteristics which, though not always exclusively exhibited by them, are much more common with them than with volcanic quakes, and *vice versa*. If in addition to close proximity to a volcano the quake has other characteristics of the volcanic class, and is destitute of the characteristics of the tectonic class the probability of its volcanic origin is increased.

We have noted in the last chapter the absence in Japan of any tendency of earthquakes to associate themselves with volcanos. Their independence of them is a very striking fact, which is strongly reinforced by the tectonic character of a majority of them. Most Japanese earthquakes originate at sea, and their vibrations on land indicate that they have travelled a considerable distance and that the disturbance which originated them must have involved a much higher aggregate of energy than is exerted even in the most forcible volcanic quakes. The great Mino-Owari quake of 1891 originated on land, and its obvious cause, clearly disclosed in the great dislocation, proves its tectonic nature. The Kumamoto and the Noto quakes, both of great power and making their records on the pendulums of Von Rebeur-Paschwitz in Germany were land quakes and of unmistakable tectonic origin. But the majority of the Japanese quakes originate on that great slope of the sea-bottom which leads down to the Tuscarora Deep. It would be going too

far, however, to assert that none of the Japanese quakes are volcanic. There is a portion of them, possibly not far from three per cent., whose origin may be volcanic.

Our knowledge of Philippine earthquakes is much less complete than that of the Japanese. Enough is known of them to indicate that they are similar in character to those of Japan. Most of them, and indeed all of the more important ones, have the tectonic character. A few of them have their epicentres on land near the eastern coast of Luzon and near the main mountain-ridge from Tayabas to Bulacan, and even as far north as Nueva Ecija. Manila, being situated within fifty to eighty miles of these epicentres, has repeatedly suffered from their vibrations. In 1863 most of its principal structures were seriously injured and some of them destroyed. In 1880 a severe shock badly damaged the cathedral, the Governor's palace, and many other masonry buildings. Shocks of slightly less intensity visited the city in 1869 and 1872, and again in 1891. The traditions of the city as transmitted by the priests and friars seem to indicate a perpetual dread of earthquakes ever since the advent of the Spanish, and a severe shaking every ten or fifteen years is looked upon as being in conformity with the due course of nature.

The situations of the principal epicentres appear, like those of Japan, to be quite independent of the volcanos, as they occur in a portion of Luzon where no recent volcanos exist. In the vicinity of the great active volcano El Mayon in Albay, or around the smouldering Volcan de Taal, there is but little seismic action, and that little may be of volcanic origin. But great destroying shocks which are felt far and

wide after the usual manner of tectonic quakes originate along the eastern slope of the island, sometimes above the sea-surface, sometimes below it. They invariably are followed by a series of after-shocks and those originating at sea are sometimes followed by great sea-waves rolled in upon the coast, of which Dr. Rudolph records several.[1] He also records shocks reported by ships at sea in the same vicinity.

According to older concepts Japan and the Philippines are merely segments of that supposed "fiery great circle," which surrounds to the Pacific from Cape Horn to Behring Sea, thence down the eastern archipelagos of Asia to the Dutch East Indies. But in reality there is no such circle. There is nothing more than a considerable number of districts of high seismicity separated by wide intervals of low seismicity. They are situated upon the most pronounced slopes leading from the land to sea-bottom. The epicentres are sometimes near the bottom of those slopes and more often than otherwise part way down. This is the case in both Japan and Luzon. Between those two fields of seismic action there is no evidence of any exceptional degree of seismicity.

Similar intervals of feeble seismicity or of none at all separate the Japanese field from those to the north-east of it. There is an inconsiderable amount of seismic action along the Kurile chain and the Kamtchatka coast and along the western Aleutian chain. It appears to increase in the eastern Aleutians. We should expect to find along this island barrier of Behring Sea indications of volcanic rather

[1] *Beiträge zur Geophysik*, vol. i., with map.

than tectonic disturbances, for active volcanos are numerous there. On the contrary, most of the well-defined quakes from that quarter are clearly of the tectonic class and have been made known to us in their character of world-shakers by the records of horizontal pendulums in Canada, Europe, and even at their antipodes in South Africa. In truth the profound depths of the ocean just off the eastern part of the Aleutian chain is one of the great breeding-grounds of world-shakers. A rather small basin in the ocean-bottom has here a depth of nearly four thousand fathoms, and the descent to it is by a long and strong gradient.[1] Earthquakes had been known, however, all along the thousands of miles of coast-line of Alaska and the Aleutians for more than a century. The Russians had often experienced them and had associated them with the volcanos around Behring Sea. Nor is it at all improbable that many of those thus noticed were the results of volcanic activity.

Between the Alaska-Aleutian field and the coast of California earthquakes are infrequent. But from Cape Mendocino southwards the seismicity increases again. Prof. E. S. Holden has made a study of the seismic history of the State of California and has catalogued the recorded quakes from 1769 to the end of 1896.[2] During the nineteenth century

[1] It is not surprising that such powerful quakes should be unfelt on land and be discovered only by instruments installed thousands of miles away. The senses are affected by earth vibrations in proportion to the acceleration of the vibrating earth particle. From what has already been explained, it will appear that at a distance of a thousand miles the acceleration must be imperceptible except in lofty buildings or under special conditions which make it exceptionally noticeable. This reduction of acceleration with distance is due as much to the increase of period as to the decrease of amplitude.

[2] *Smithsonian Miscellaneous Collections*, vol. xxxvii.

this catalogue shows ten quakes whose intensity must be classed as high as No. 8, R. F., four of them as No. 9, and one of them as No. 10. The latter was the Owen's Valley or Inyo quake of March 26, 1872, which has been referred to already. In the same period were twenty-nine quakes which might be classed as No. 7, at least—some of them perhaps as No. 8—while lighter shocks go into the hundreds. From 1850 to 1886 inclusive the catalogue shows 254 in San Francisco, and 514 in the State of California, exclusive of those of San Francisco. Shocks also have been reported as felt by vessels at sea off the coast near Cape Mendocino, but not often from other points.

The California quakes as a class suggest a tectonic origin. High intensities are not common. The lighter intensities are felt over considerable areas, which suggest great depth of focus. The seismographic traces show considerable length of period and well-marked separation between the short preliminary tremors and longer waves, which is indicative of considerable distance travelled by the vibrations between the centrum and the recording station. The deep foci, the long periods, the absence of small tremors, the considerable areas over which light vibrations are felt, are indicative of tectonic rather than volcanic origin.

Proceeding southward along the Pacific coast no marked development of seismic activity appears until we reach southern Mexico in the state of Colima. Occasional reports are made of quakes in the vicinity of the great volcano Colima, and the City of Mexico has been visited by a few shocks of moderate intenisty and by occasional light tremors. But the developments of seismic action are not

great. So, too, in the vicinity of the great mass of Orizaba, earthquakes are occasionally felt, and some of them have been somewhat forcible, but they do not appear to have been frequent or destructive. In the state of Oaxaca, however, there is a sudden and great increase of seismicity. It occurs in a long and rather narrow belt extending parallel to the coast reaching into Guatemala and through the entire length of the Central American states to the Isthmus of Panama.

The Central American and south Mexican seismic regions are strongly contrasted with those we have thus far mentioned, as they indicate a volcanic instead of a tectonic origin. These regions are famous for their volcanos, which are more numerous and closely adjacent than anywhere else in the world. The quakes themselves are so closely associated with volcanos as to leave no doubt as to their origin.

They have been the subject of careful research by De Montessus de Ballore, who was from 1881 to 1885 military instructor in Salvador, and took advantage of the opportunity to examine old records of the Central American states. These he afterwards supplemented by researches in Europe.[1] The seismic distribution in the Central American district is so typically volcanic, the interdependence of earthquakes and volcanos is there so close and so obvious, and contrasts so conspicuously in this respect with what we observe in regions of tectonic quakes that it seems advisable to enter a little more fully into a description of details.

The volcanos of Central America form a linear series ex-

[1] *Tremblements de Terre et Eruptions Volcaniques au Centre-Amerique depuis la Conquete Espagnole jusqu' à nos jours*, Dijon, 1888.

tending from Panama to Oaxaca. This series is readily subdivided into six portions, separated by narrow intervals in which no recent volcanic action exists. The resumption of the volcanic chain at each of the intervals of repose is not always in the exact prolongation of the original line, but sometimes, as in Nicaragua, or Guatemala, by an *echelon*. The volcanic line is always south of the main divide, or Sierra Madre, and very near the Pacific coast. In fact, the Pacific Ocean is in full view and near at hand from the summit of every important volcanic cone, and the cones themselves are conspicuous objects from the decks of steamers passing up the coast on their way from Panama to Acapulco. Along these lines between Panama and Oaxaca there are 140 volcanos, of which thirty-one are known to have been in eruption since the advent of the Spaniards in 1526.

The first of the six subdivisions begins in Chiriqui in the state of Panama, and extends westward through Costa Rica close to the Pacific coast as far as the volcano Orosi, on the Nicaraguan border. In this subdivision are the large volcanic piles Irazul, Turrialba, and Poas overlooking the cities San José and Cartago. All have been active within the historic period. Cartago has been destroyed by earthquakes no less than four times within the last two centuries, and each time the catastrophe was associated with the re-awakened activity of Irazul. The westernmost pile of this subdivision, Orosi, is reported to have been in eruption in 1827.

The second subdivision begins in Lake Nicaragua, about fifteen miles north of Orosi *en echelon*. Two large cones rise out of the lake, Madera to the east, and Ometepe near

the south-western shore. Madera has not, it is believed, erupted within the historic epoch, but Ometepe has done so several times. The line continues to Mombacho, hard by the city of Grenada, through Masaya, Momotombo, Las Pilas near Leon, and ends at Coseguina at the entrance of the Bay of Fonseca. All these may be regarded as active volcanos which have erupted several times, Masaya many times, within the historic period. Las Pilas is a new volcano which broke out as an entirely new creation in 1850. The eruption of Coseguina in 1835 was a convulsion of a tremendous order, and ranks as one of the most memorable occurrences in the volcanic history of the nineteenth century.

The third subdivision begins across the Bay of Fonseca at the active vent Conchagua, and extends through the volcanos San Miguel, San Vincente, San Jacinto, Santa Ama, Ilopango, and Izalco. The last two are remarkable as newly formed volcanos, a creation which vulcanologists view with as much interest as astronomers do the appearance of a new star in the heavens. Three new volcanos have been formed in Central America within one hundred and fifty years: Izalco, February 23, 1770; Las Pilas, April 11, 1850; and Ilopango, December 20, 1879.

The fourth line begins in Guatemala, and, on the whole, is the most impressive of all. It includes the very large and intensely active cones Pacaya, Fuego, and Santa Maria. Westward the remaining subdivisions continue with a slowly decreasing energy through the Mexican states Chiapas, Tehuantepec, Oaxaca, and even as far as Acapulco, in the state of Guerrero.

Throughout this fifteen hundred miles of volcanic coast-line earthquakes have always been abundant and often highly destructive. From the scanty records preserved, De Montessus has succeeded in bringing to light 772, and most of them are of a degree of energy sufficient to cause a memorable amount of destruction, or to leave a deep impress upon the public mind. One of the most striking features is the fact that the epicentres of all quakes occur in close proximity to the volcanos and never at a distance of more than four or five miles from them. In the older country, the Sierra Madre, the main continental region lying back of the volcanic belt, it is not known that earthquakes ever originate. The association of the earthquakes with the volcanos is so intimate and so uniform that no question as to their interdependence seems possible. Nor does there seem to have been any notable quake which suggests a tectonic origin. For all of them indicate shallow centra. Even those of high intensity in their epifocal tracts quickly lose it as they spread out, showing that rapid diminution of energy away from their epicentres which has already been described in the case of the Casamicciola earthquake.

Entering South America at Panama two seismic regions are before us, both of them, however, at a great distance: one in Venezuela, far to the eastward, the other in Ecuador, far to the southward. Our knowledge of the earthquakes in the Andes is usually too insufficient in detail to enable us to form a very definite opinion as to their nature and class. Some of them originate in places which are not far distant from volcanos, but whether near enough to create

the presumption that they are really associated with them is not always known. Boussingault and Humboldt both remark upon the fact that epicentres are very often far away from any volcano. On the other hand, some of them are certainly so near a volcano and occur so concordantly with its eruptive activity that no doubt can remain. On the other hand, many Andean quakes display the tectonic character so clearly that their origin is equally free from doubt. That both classes of earthquakes are abundant and intermingled in the Andean regions and along the adjoining coast seems very probable, though want of detailed knowledge of their characteristics prevents us from forming an estimate of the relative importance of each class.

From a point about two hundred and fifty miles north of Callao to beyond Valdivia, in Chili, the South American coast has been subject to earthquakes of the greatest energy. Sometimes they have originated upon land, sometimes upon the sea-bottom. In the latter case great sea-waves have been rolled in upon the coast with disastrous effect. The terrible wave that overwhelmed Callao in 1867 will rank with the deluges which destroyed Arica in 1868, and Simoda in 1854. The provinces of Ica and Arequipa, in Southern Peru, have been celebrated for their destructive quakes. It is the Chilian coast and Andes, however, in which the seismicity of South America reaches its highest development. Extending through more than twenty degrees of latitude, from a hundred miles north of Arica to a hundred miles south of Valdivia, it embraces several seismic districts. It may be considered an open question by some whether this great belt should be considered as a single seismic region or

treated as three or more separate regions having no interdependence. Decided preference is here given to the latter view.

The most notable of the several districts, whether they be regarded as interdependent or not, is the southern one occupying the sea-bottom between the Island of San Juan Fernandez and the Chilian coast between Lat. 26 and 39. This district is remarkable for the number of quakes of the highest order of magnitude which have originated there. The long dislocations described by Humboldt and Darwin have already been alluded to.

To complete our glance at the circuit of the Pacific Ocean we may look for a moment at the seismic district of New Zealand. Australasia is but little troubled by earthquakes, and the only portion of it in which they are at all noteworthy is in Cook's Straits, separating the two principal islands of New Zealand. Mr. George Hogben, of Timaru, N. Z., has devoted much time to collating historic accounts of Australasian quakes, and during the forty-three years preceding 1891 secured about eight hundred for New Zealand alone,[1] the number occurring elsewhere being remarkably small. The subject was taken up in the Australian Association for Advancement of Science in 1891, and a committee similar to that of the British Association was formed, with Mr. Hogben as secretary, to co-operate with the British committee in collating information and observing earthquakes. The reports of the association show that for several years the number of quakes in New Zealand averaged about fifty per year, the intensities ranging from 3 to 7 R. F.

[1] *New Zealand Jour. Science*, Nov., 1891.

The only destructive ones recorded in the recent history of the islands were those of 1848, described by Lyell in his *Principles of Geology*,[1] and one of somewhat less power in 1855.

The seismicity of Australia is on the whole very low, and even in Cook's Straits, where it is highest, it falls much below the districts of pronounced seismicity in Japan and Italy. Nevertheless, it is an instructive locality, as it includes quakes of both classes. Those in Cook's Straits or its immediate vicinity are clearly tectonic, the occurrence of 1848 described by Lyell being visibly connected with and immediately resulting from a marked tectonic movement. The volcanic quakes are as clearly associated with the eruptive action in the group of volcanos around Mts. Edgcombe and Tarawera.

In this very cursory view we have taken of the seismic regions surrounding the Pacific Ocean, we observe that they consist of a number of widely separated areas, the intervals between them being areas of repose seldom disturbed seriously by earthquakes. A portion of the disturbed areas is characterised by quakes of volcanic origin, while the greater portion is apparently tectonic. But the most suggestive feature of the distribution appears when we consider it in relation to De Montessus de Ballore's deduction as to the situation of earthquakes. He finds that in most cases they occur where the variations of topographic relief are greatest.[2] As no one is qualified to speak with so much authority on

[1] Also in *Westminster Review*, 1849.

[2] "Relation entre le Relief et la Sismicité," *Comptes Rendus*, cxx., 1895, pp. 1183–1187.

the subject of seismic geography it is well to quote his views in some detail. He says:

1. In a group of adjacent seismic regions the most unstable (*i. e.*, most subject to be shaken by earthquakes) are those which present the strongest differences of relief, *i. e.*, the most pronounced general slopes. 2. The unstable regions are associated with the great lines of corrugation of the terrestrial crust, either emerged or submerged.

These laws are applicable with the qualification that absolute seismicity is not strictly proportional to the slope. This is due to the fact that instability must necessarily be dependent in some measure upon the nature of the ground. It is probable, moreover, that other factors than the relief are involved in the problem. In other words, the two laws just stated express necessary conditions, but not all of them. There are doubtless others to be discovered. Whatever they may be, we may reckon at only ten per cent. the exception to the law of relief. The percentage of exceptions is greater in laws of details of which the following is a summary.

a. Mountain districts are generally more unstable than plains. But not all mountains are unstable nor are all plains stable. In this relation the centres of disturbance have a distinct tendency to group themselves at the foot of the highlands when the slope varies suddenly. There is evidently a line of least resistance there.

b. The short and steep flank of a chain is the most unstable. This law has exceptions only when the longer slope is at the same time the roughest and most diversified, though presenting an easier average slope.

3. Rapidly deepening littorals, especially if they border important mountain ranges, are unstable, while gently sloping littorals are stable, especially if they are the continuations of flat or slightly diversified regions.

These generalisations of De Montessus are on the whole fairly sustained by the facts presented by the seismic districts bordering on the Pacific. While there is not a strict proportion between relief and seismicity, there is in general and with very little exception a markedly greater frequency and energy of seismic action upon the longer and steeper slopes than upon the gentler ones. As this law appears to be general it will be of some interest to look briefly at its manifestations elsewhere.

The southern and south-western fronts of that long chain of islands extending from the north-western extremity of Sumatra through the Dutch East Indies as far as New Guinea, are a region of both volcanos and earthquakes. Toward the Indian Ocean the littoral along many parts of that front descends rapidly to a depth of nearly four thousand fathoms. Upon this littoral many earthquakes of the world-shaking class have originated. To speak only of those occurring within the last few years, there was the great Ceram occurrence of October 12, 1899, the Java quake of September 30, 1899, and the Sumatran quake of 1895, which fell but little below the great Bengal-Assam convulsion in power.

Throughout that vast archipelago which lies between the Philippines and this chain of islands earthquakes are common, and in the Celebes particularly they occur with much more than usual frequency. The sea-bottom throughout

this archipelago varies extremely in depth, having many shallows and reefs, with deep passages and basins giving a highly diversified profile and that relief which De Montessus associates with seismicity.

We find a similar relation between relief and seismicity in the long seismic belt which begins in Venezuela and extends into the Windward and Leeward Islands of the West Indies and through San Domingo and Jamaica. Caraccas, the "City of Earthquakes,"[1] is situated among the mountains of one of the great spurs of the Andean system, and the descent is rapid to the coast at La Guayra and thence to the depths of the Caribbean. Throughout the Windward and Leeward Islands the frequency is high, though the intensity is seldom great. The earthquake of 1867 at St. Thomas, with its great sea-wave, however, will be recalled in this connection, and also the fact that just north of this island is the deepest abyss of the Atlantic Ocean within a distance of about forty miles. San Domingo, Jamaica, the eastern part of Cuba, and Porto Rico are also frequently shaken, and their seismicity may be put into relation with the fact that the sea-bottom in the vicinity of those islands is one of the most rugged and highly diversified in its profiles of any part of the earth.

[1] An interesting account of the earthquakes of Caraccas is given by Mr. H. D. Warner in the *Atlantic Monthly* of March, 1883, in an article entitled "A City of Earthquakes."

CHAPTER XVI

SEAQUAKES

The Sea has its Quakes as Well as the Land—Difficulty of Securing Well-Observed Data—Dr. Rudolph's Investigations of Seaquakes—Description of the Quake on the Water—Varying Intensity of Seaquakes—Sounds Heard in Seaquakes—Explanation of Tremors in the Water, which can Transmit only Normal Waves—Energy of Sound-Waves in Water—Other Disturbances than Normal Vibrations are Rare—A Few Instances of Extraordinary Agitation of the Water Not Attributable to Normal Vibration—Submarine Seismic Regions in the Atlantic—St. Paul's Rocks—The Equatorial District—Submarine Quakes near the Azores—In the West India Deep—Sea-Waves of Seismic Origin—Those of the Peruvian-Chilian Coasts—The Great Arica Quake and Sea-Wave of August 13, 1868—Its Record on the Tide Gauges of Japan, Australia, and California—Its Speed of Propagation—Causes of Such Waves—Sudden Uprise and Downfall of Considerable Areas of the Ocean-Bed—Downfalls More Frequent—The Simoda Wave of December 29, 1854—Bay of Bengal, December 31, 1881—Dr. Rudolph's Views of the Causes—Krakatoa Wave

THE sea has its quakes as well as the land. Our opportunities for recognising and studying the two categories are, however, widely contrasted. By means of delicately poised instruments we can detect and measure the minutest amplitudes and the slowest and gentlest periods of the land-quake, while on the water the use of such instruments is impossible. The seaquake can be detected only when the water is put into a state of vibration sufficiently energetic to cause the ship and its loose objects to tremble and thus

affect the senses. The vast waves rolled in upon the shores of continents are much too flat and too slow in period to be perceptible to the senses of the traveller on shipboard. But sometimes the vibration of the water imparted to the ship is sharp enough to arouse the attention of passengers and crew. Sometimes it is forcible enough to cause alarm or terror. In very rare cases it causes utter dismay and the feeling that destruction is at hand.

Dr. Emil Rudolph has devoted much time and labour to the research of reports from ships' logs and other sources bearing upon the subject of quakes observed at sea.[1] From the very large number of reports he has brought together we may select a few which will illustrate the kind of sensations experienced by the officers and crews of vessels during a seaquake.

Captain Gales, of the ship *Florence Nightingale*, reports that

"on January 25, 1859, while in N. Lat. 0° 48′ and W. Long. 29° 16′, St. Paul's Rocks being about ten miles N. W. by N. of us, we felt a strong shock of an earthquake. It began with a rumbling sound like distant thunder and lasted about forty seconds. I was quite well acquainted with earthquakes, as I had experienced a good many on the west coast of America, but never had I felt so severe a one. Glasses and dishes rattled so vigorously that I was surprised to find them uninjured. A good many objects fell down and it was as if the ship were grinding upon a reef. At once arose from all sides the cry, 'The ship has struck!' and the watch came tumbling up in hot haste. I was much alarmed and looked over the side of the ship in order to see the reef, but quickly formed my opinion and quieted the commotion by the

[1] These researches are embodied in two long articles in *Beiträge zur Geophysik*, vols. i. and ii., and form very interesting and instructive reading.

explanation that it was only an earthquake."— Petermann's *Geogr. Mittheil.*, xv., 1869, p. 97.

Another report from a locality not far from the above is the following: "On January 28, 1883, in N. Lat. 1° 38′, W. Long. 27° 40′, in clear weather and a light sea, suddenly we heard, about 7.47 P.M., a strange submarine noise not unlike distant thunder or still more like the distant firing of heavy guns. At the same time there was a vibration of the ship as though the anchor had been let go, or as if one were standing on the after-deck of a screw steamer. The entire phenomenon lasted about a minute. A peculiar sensation came upon everybody as if electrified. The crew thought there must be a large stick of timber rubbing alongside. The lookout thought that the ship had struck bottom."

The foregoing are representative of the large majority of the reports of seaquakes. The ship quivers, vibrates; loose objects chatter and tremble. There is a strange noise in the sea like distant thunder or distant artillery. The first impression is as if the ship were grinding upon the bottom, and there is an instinctive rush of the crew to the deck and the bulwarks to see if the ship is not aground or on a reef. But the situation is soon recognised. The ship is seen to move steadily onward with unchecked speed, she rises and falls to the swell of the sea without shock, the water is dark and fathomless. The tremor soon passes and the nature of the phenomenon is at length apparent.

Although the trembling of the ship and the strange roar from the sea are the most common and exclusive indications of the seaquake, there occur more forcible indications in a few instances. As might be expected there are degrees of

intensity in seaquakes just as there are in landquakes, though the means and agencies by which they are made sensible are much more limited. Among many hundreds of reports from ships at sea which Dr. Rudolph has collected are a few which indicate intensities of a high order. Thus one master of a vessel reports: "We felt a shock so strong that the entire crew was brought to its feet at once; the wheel flew from the hand of the steersman and I myself was flung down upon the deck." He quotes Virlet d'Aoust, a French geologist, who in a paper on earthquakes states that in an earthquake experienced on the coast of Asia Minor: "Our ship was over the epicentre and was so severely shaken that at first the Admiral feared the complete destruction of the corvette." Heavy objects including cannon and their carriages were thrown up from the deck. The corvette itself seemed to be hurled upwards. The statement that heavy objects have been lifted from the deck and the vessel itself lifted as if projected upwards is by no means unique, for Dr. Rudolph has collected a considerable number of them. The exact amount of credence we ought to concede them or the precise interpretation we ought to give them is another matter. We seem to be justified in believing that in rare cases the power of the shocks may be great enough to render standing on the deck as difficult as it sometimes is on land. It may even be great enough to cause the fear that the vessel is being shaken to pieces.

The tremors imparted to the vessel from the water and the strange sounds from ocean depths are readily explained. The only form of elastic wave which a fluid medium can transmit is the normal wave. This mode of vibration the

water of the ocean receives from the materials which constitute its bed. These vibrations transmitted through the earth may, before reaching the under-surface of the water, be either normal or transverse. A transverse wave passing through a solid elastic medium and reaching a bounding surface which separates it from a different solid elastic medium is split up into four portions: 1, a transmitted transverse wave passing into the new medium; 2, a transmitted normal wave; 3, a reflected transverse, and 4, a reflected normal wave. Also a normal wave reaching a similar boundary is split into four portions: a transmitted normal, a transmitted transverse, a reflected normal, and a reflected transverse wave. But there are exceptions depending upon the angle of incidence, *i. e.*, the angle at which a ray of the original wave meets the bounding surface. When this angle is less than a certain critical value there is no transmitted wave, but a total reflection.

When the bounding surface separates a solid from a fluid medium only the normal component of the wave can pass from the former to the latter. But it may have been the component of either an original normal or of an original transverse wave.

When it has once entered the water it is in a sensibly perfect, *i. e.*, perfectly elastic, medium, and is propagated through it like any ordinary sound-wave in water or in air. If the vibration is continuous, it is maintained as a continuous tremor in the water like the continuous roar of a fog-horn in air or the hum of the locomotive. But, as we have already had occasion to note, the vibrations of earthquakes are innumerable and of many periods, and all are simul-

taneous and superposed. Those whose periods are within the limits of audible sound can be heard, but as a confused murmur, due to the superposition of innumerable periods.

The energy involved in sound-waves in water far exceeds, ordinarily, the energy of sound-waves in air. This is because the energy, which is proportional to the mass of the water particle multiplied by the square of its vibratory speed is many times greater than the correlative energy of the air particle. Moreover, the elasticity of water is many times greater than that of air. It is not surprising, therefore, that the quicker vibrations of the earthquake, imparting normal vibrations to the water, should supply them with energy enough to make a ship tremble. If exceptionally strong they may, by cumulative vibration, upset loose objects and cause grave alarm. But that they can ever, under any circumstances, become energetic enough to lift the ship, or throw up heavy objects on the deck, is doubtless an illusion.

Besides this trembling of ships, sometimes taking a vigorous and most alarming form, and the roaring of the sea due to the normal waves rushing through it from the earth beneath, there are other phenomena whose accounts Dr. Rudolph has collated and which are of interest in this connection. Very rarely do the reports of seismic occurrences at sea speak of any disturbance of the water other than that due to the winds and waves. Usually they state distinctly that there was no other. But a few reports have been gathered where the agitation of the water is spoken of as being marked or even extreme. The descriptions seem to

imply nothing less than submarine volcanic eruptions of great power, and vessels sailing over the spot appear to have been caught in the swirl of the waters sometimes with disastrous results. These occurrences are extremely rare. But with the tens of thousands of ships ever sailing the seas it is well within the limits not only of possibility but even of probability that such a mishap should at long intervals

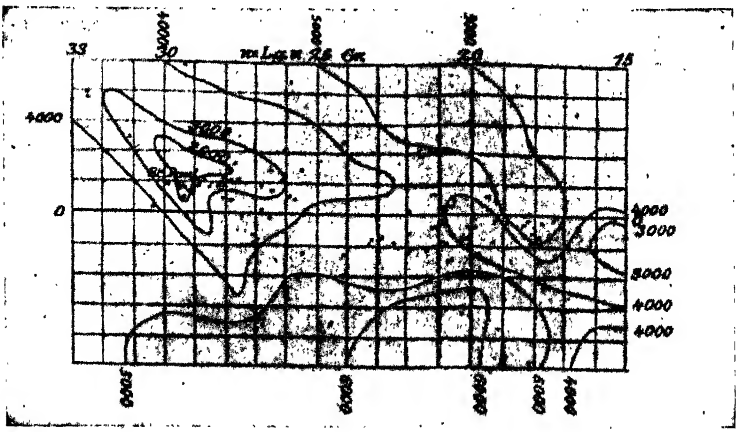


FIG. 63. Dr. Rudolph's Contours of Ocean-Bed in the St. Paul's Rocks and Equatorial Seismic Districts.

occur. Among the most valuable results of Dr. Rudolph's researches is the establishment of numerous submarine districts or regions which are seismically sensitive and give origin to an unusual number of seaquakes. Among the first of these to be determined were two, located in the Atlantic Ocean very near the equator, nearly midway between Cape Palmas on the south-eastern coast of Liberia and Cape St. Roque, Brazil. Near Lat. 1° N. and Long. 30°

W. is the volcanic reef known as St. Paul's Rocks. It is the peak, almost awash, of a submarine ridge planted upon a plateau in the ocean-bottom. On all sides the sea-bed descends rapidly into deep water. In the vicinity of these rocks an extraordinary number of seaquakes has been felt in passing vessels. Farther east, on both sides of the equator and on both sides of the meridian of 20° W., is another region from which an unusual number of seaquakes has been reported. Dr. Rudolph has charted them and put them into relation with bathymetric curves obtained from deep-sea soundings. He indicates for apparently sound reasons two entirely distinct districts separated by a wide surface interval which has furnished no reports of quakes, and separated in the depths by a strongly marked and rugged topography. To the western field he gives the name of the St. Paul's Rocks district and to the eastern the name of the Equatorial district.

Between 1845 and 1893 there are thirty-seven quakes cited by Dr. Rudolph from ships in the neighbourhood of St. Paul's Rocks, and forty-nine from the Equatorial district between 1747 and 1890. A considerable number in the latter district, however, are reported by two or more vessels which felt the tremors simultaneously from the same disturbances. The degree of seismicity thus indicated for both these places will upon full consideration be regarded as a very high one. It will occur to us that these are merely the reports of passing vessels, which chanced to have fallen into the records of navigation bureaus, while the experiences of other vessels which may have felt the same or other shocks failed to be preserved, or, if anywhere pre-

served, failed to be discovered. It is not probable that any such tremors were noted at all unless under the following conditions: 1st, they must be of a rather high degree of intensity and derived from quakes of very much more than average power to be felt under any circumstances; 2nd, they must have occurred only when the ocean was very quiet, and would hardly have been felt in rough weather. The equator, it is true, is ordinarily a belt of calms; but, on the other hand, the ocean is never still. All things considered, the probability of any quake other than one of considerable power being felt by a passing vessel at sea is a very small one, and the probability of its record finding a resting-place where it may in after-years reach the eye of some investigator is also a very small one. Thus the number of quakes recorded in these two oceanic regions must be but a minute fraction of those which have probably occurred there.

Another district from which seaquakes have been reported with exceptional frequency in the North Atlantic is the neighbourhood of the Azores. Between these islands and the coast of Portugal it may be remembered that the great quake originated which on November 1, 1755, destroyed Lisbon, though probably its epicentre was much nearer the continental shore than the islands. Although seaquakes are common anywhere along this belt of sea-bottom (between the Azores and Portugal) they appear to cluster much more thickly near the islands or near the mainland than midway between them. Whether this is due to the fact that vessels which are likely to observe and report them pass much more frequently near the two localities than

between them, or is due to a real difference in the number of quakes, is uncertain.

The West India Deep, that profound basin of the Atlantic lying north of the Lesser Antilles and east of the Bahamas, where the Atlantic has its greatest depths and where its bottom has its greatest inequalities and reliefs, is another district from which an unusual number of seaquakes has been reported.

The most impressive accompaniments of seaquakes are those gigantic waves in the ocean which are generated at the same time and doubtless by the same kinetic causes as the seaquakes themselves, and which break upon adjacent coasts with disastrous results to cities and their inhabitants. These have been known through a long period of history in the eastern Mediterranean, where they have ravaged the coasts of Syria and Asia Minor, as well as the shores and islands of the Ægean. It is not a little remarkable that the introduction of the great work of Suess, *Das Antlitz der Erde*, is an essay which attributes the Noachian Deluge to a legend of a mighty sea-wave rolled in upon the lands of Chaldea from the Persian Gulf. Although western Europe and the eastern coasts of North America have had little experience of such visitations, the dreadful memory of the destruction of Lisbon still survives.[1] In the West Indies, though happily not common, the history of the last four

[1] The literature of the Lisbon disaster is very copious and might form a library of itself. The best succinct account of it that occurs to me is given by Lyell in his *Principles of Geology*. There is also an interesting paper in vol. xii., *Trans. Seism. Soc. Japan*, 1888, communicated by E. J. Pereira, being an abstract of a rare old pamphlet published in Lisbon in 1756 by an eye-witness a few months after the catastrophe.

centuries has preserved accounts of several of them. But, generally speaking, the Atlantic both north and south of the equator has been remarkably exempt from them.

It is off the Pacific coast of South America from the equator to 45° of south latitude that these waves originate with greatest frequency and also in greatest power. Especially in the vicinity of the angle where the Peruvian and Chilian coasts meet have they been most numerous and formidable. The harbours of Pisco, Arica, Tacna, Iquiqui, and Pisagua have been repeatedly subject to these destructive invasions. Usually they are preceded by a violent earthquake, and the inhabitants, taking due warning therefrom, betake themselves to the hills. The sea-wave, however, does not always follow the earthquake; in fact, in a great majority of cases, it does not. But it appears often enough to arouse serious apprehension of its coming whenever the ground is strongly shaken. The first indication of the coming disaster is the withdrawal of the sea from the shore, leaving bare the bottom of the harbour. A few minutes later the sea returns in a high wave of resistless power which overflows the adjoining flats. Again it withdraws and still again returns. The oscillations may continue through many hours and even for two or three days at intervals, *i. e.*, periods of about fifteen minutes, and with slowly diminishing amplitudes (amount of rise and fall) until they gradually die out.

The most memorable seaquake of this locality occurred August 13, 1868. The coast of South America was shaken all the way from Guayaquil in Ecuador to Valdivia in Chili, the highest intensity being manifested in the neighbourhood

of Arica.[1] The force of the quake in this town was very great, throwing down most of the structures and producing landslips. A few minutes later — precisely how many minutes is not known—the sea was observed to retire slowly from the shore so that ships anchored in seven fathoms of water were left high and dry. A few minutes later still it was seen returning in a great wall or "bore," which caught up the ships in the roadstead and swept them inland as if they were mere chips of wood. Among them was the U. S. steamer *Wateree*, one of the improvised war-vessels of the blockading fleet of the Civil War, which was carried inland nearly half a mile and left with little injury on shore by the recession of the wave.

The wave in the ocean generated in this quake made itself felt on the coasts of Australasia, Japan, Kamtchatka, Alaska, Oregon, and California. In the harbour of Hakodate, in Japan, a series of waves was registered upon the tide gauge. The ordinary tide at that port is about 2½ to 3 feet. On this occasion the water rose and fell ten feet (double amplitude) with a period of about twenty minutes for a complete oscillation. It had taken the first wave twenty-five hours to traverse the distance of about 7600 nautical miles. On May 9, 1877, another seaquake of similar magnitude had its origin in the same neighbourhood. It is known as the Iquiqui quake. A vast sea-wave invaded with disastrous effect the towns of northern Chili and southern Peru, repeating the havoc of 1868. At Arica

[1] A very full account of this great seaquake is given by F. von Hochstetter in *Sitz. der K. K. Akad. der Wiss.*, bd. lviii., 1868, ii. abth., and it is summarised in Petermann's *Geogr. Mitt.*, 1869.

the stranded hulk of the *Wateree* was picked up and swept farther inland. Like its predecessors, this wave was felt all over the Pacific. At Samoa the height of the waves varied from six to twelve feet; in New Zealand and Australia from three to twenty feet; in Japan, from five to ten feet.

It should be borne in mind, however, that as a sea-wave progresses from deep water into gradually shoaling water it increases its height and the steepness of its front slope until it begins to comb or break. The height measured by the tide gauge in a seaport is therefore no measure of the height of the wave far out in the ocean, and, in fact, the mid-ocean height of the wave is doubtless measurable in inches while the inshore height is measurable in feet.

Since a wave must travel its own wave-length in the time indicated by its period, we shall have no difficulty in computing the lengths of these giant waves as soon as we know their periods and speeds of propagation. If it requires twenty-five hours for a wave to travel from Arica to Hakodate, a distance of 7600 sea-miles, we have a speed of about three hundred miles per hour, or five miles per minute; and if the period as shown on the tide gauge at Hakodate is twenty minutes, we have a wave-length of one hundred nautical miles. We have no means of knowing the height of the wave in deep water. We can only say that it must be much less than on a neighbouring shore. Obviously it is greater near the origin than away from it.

Since the speed of propagation of water waves is dependent upon the depth of the water, much computation has been devoted in times past to the problem of deducing the depths of the ocean from the speeds of sea-waves. The

measurement of oceanic depths by actual soundings has been one of the results of the demands of ocean telegraphy, and it has been prosecuted so vigorously that we are rapidly accumulating direct and positive knowledge on this point, and the computation of depths from the speeds of sea-waves has now become an academic problem rather than a practical one.

There has also been much discussion of the specific modes of action which have originated these waves. It must be admitted that the varying opinions or views on this subject appear to have been influenced often by preconceived views of geodynamic action, and the discussions have in some instances shown a tendency to mould facts to fit hypotheses; to belittle or explain away facts which may be adverse to such preconceptions and unduly magnify those which may seem to favour them.

There are, however, several facts or groups of facts which admit of no dispute or qualification. (1) A great sea-wave of this kind implies that somewhere in the depths of the ocean, usually not very far from shore (say within one or two hundred miles of it), a displacement of the under-surface of the water occurs. It may be a sudden upri-ing of the sea-bottom, lifting the overlying mass of water, or it may be the sudden dropping of the bottom, carrying the water down with it, or it may be a sudden volcanic outbreak discharging lavas with their invariably associated contents of intensely hot steam or other occluded gases into the overlying ocean with sufficient expansive force to lift the water. In any event there must be some lifting or depressing force acting at the sea-bottom or on the littoral.

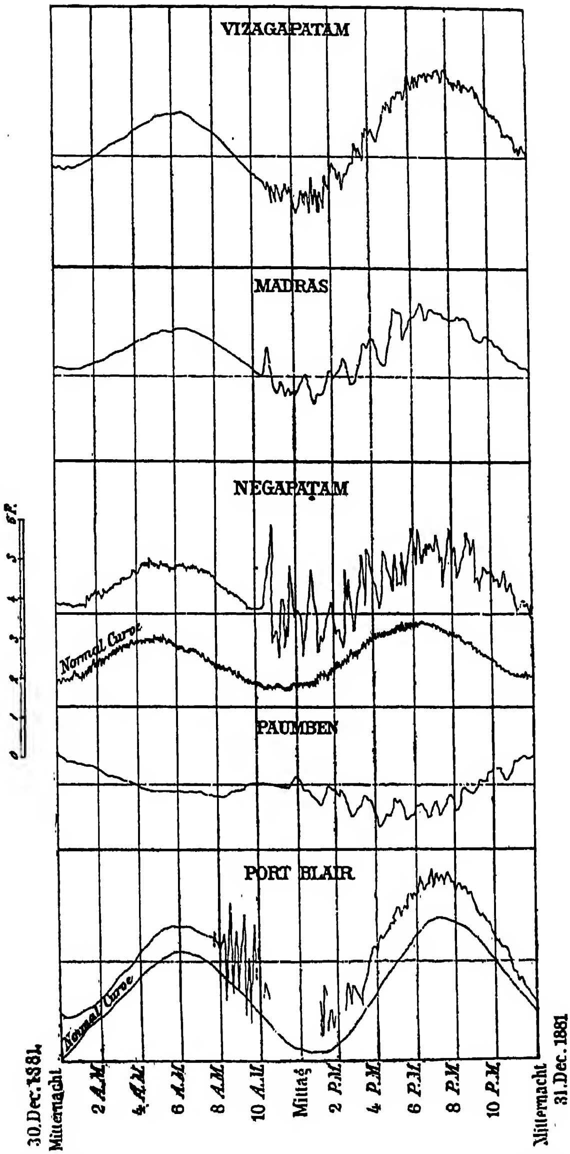
(2) The energy involved in the movement must be great enough not only to lift an unknown thickness of sea-bottom, but the overlying water several thousand metres in depth. The area of displacement must also be a very large one, *i. e.*, hundreds and perhaps thousands of square kilometres in extent.

(3) The movement must be sudden, *i. e.*, rarely taking so much as a minute in its execution, and more commonly taking but a few seconds.

By putting these three conditions into relation with known facts of observation we may indicate the probable causes of such waves. In most of the seismic sea-waves the first motion perceived on shore is the withdrawal of the water. In a few instances the reverse occurred. The water was first perceived to rise. The latter appears to have been the case in the great wave which overwhelmed Simoda in Japan December 29, 1854.[1] One of the instances in which the tide gauges showed first a rising wave occurred in the Bay of Bengal on December 31, 1881. The epicentre of this quake appears to have been about 550 miles east of Madras and strongly affected all the tide gauges of the surrounding bay. In every instance the trace of the gauge first shows a sharply rising wave. Another powerful quake in the Moluccas (November 26, 1852), was accompanied by sea-waves whose rising branches were first felt in the harbours of Banda-

[1] According to the report of Vice-Admiral Putiatin on board the Russian frigate *Diana*, which was lying in Simoda harbour at the time, very powerful shocks of an earthquake were felt at 9.15 A.M. At a little before 10 o'clock a huge wave rolled into the bay and quickly overflowed the city. From the above interval of time and the known speed of propagation it may be inferred that the epicentre was over 100 miles from shore and out in the Tuscarora Deep in 4000 fathoms of water.

PLATE X.



Tide-Gauge Records of the Seaquake in the Bay of Bengal.

Neira, Great Banda, Amboyna, Ternate, and Ceram. It was so in the Bay of Manila, which was invaded by a wave on June 3, 1863, the rising branch being first felt.

On the other hand, the great sea-waves starting from the Peruvian and Chilian coasts or littorals, so far as existing accounts enable us to judge, first manifest themselves almost if not quite invariably as falling waves. The same is true in most cases of the sea-waves in the Mediterranean. The wave which caught the U. S. steamship *Monongahela* on November 18, 1867, and flung her into the streets of St. Thomas, Danish W. I., began as a falling wave. Most of the sea-waves which break upon the eastern coasts of the Philippines, Japan, and Kamtchatka begin as falling waves.

When the first motion is a falling one we cannot avoid concluding that a hollow or depression of the ocean surface has been formed (presumably a sudden drop of the bottom) towards which the surrounding waters flow from all directions. Meeting at a common centre the depression is filled and the inertia of the water piles up a wave above the original mean level. Its recoil generates a secondary wave and the oscillations repeat themselves until the energy of the vibration is absorbed. If instead of a sudden drop of the bottom there be a sudden upheaval, the general nature of the sea-wave is the same, but the phase alone is reversed.

There seems to have been a tendency in recent years on the part of some geologists to contest the idea that sudden upheavals or downthrows of large areas of the earth's surface can occur. The old classic illustrations of Lyell taken from the Chilian coast, from the delta of the Indus, and from Cook's Straits, have been disputed and testimony to the

contrary been arrayed against conclusions which had long been accepted. But as if to rebuke these disciples of little faith a series of events has occurred within the last twenty years which are strongly confirmatory of the general view advanced by Lyell. The great Mino-Owari dislocation in Japan, the Bengal-Assam dislocation in 1857, the similar event in Sumatra in 1899, the Bavispe fault of 1887, were all occurrences of the same order of magnitude and energy involving the same broad principle of causation. If any one of them had occurred in the deep sea they would assuredly have generated a vast sea-wave sufficient in magnitude to have crossed the Pacific and impressed great disturbances on tide gauges eight thousand miles away. It appears to be a case where conjectures must give way before a series of observed facts, not merely once but many times repeated.

Dr. Rudolph appears to be impressed with the conviction that the most satisfactory explanation of these waves is to be found in submarine volcanic eruptions. With great ability and acumen he has discussed in considerable detail the wave generated in the memorable Krakatoa eruption in 1883. His argument is very convincing so far as it applies to this single case. But the Krakatoa eruption was unique. It was the most energetic of which we have any detailed account. It was more like an explosion than like what is usually conceived of as a volcanic eruption. It occurred in comparatively shallow water. If the water had been four or five miles in depth it is questionable whether any great wave would have been generated notwithstanding the immense energy involved. But even conceding that this unprecedented explosion might have been adequate to lift a

layer of water four or five miles in depth it would still remain a most exceptional occurrence. Volcanos as we know them pour forth their lavas and vapours slowly. That an eruption should have occurred of such stupendous power and with such suddenness as to generate a considerable sea-wave capable of crossing the Indian Ocean is a remarkable circumstance. It is believed to be the only one of the kind of which history has preserved any authentic record.

APPENDIX

DE MONTESSUS DE BALLORE'S TABLES OF THE DISTRIBUTION OF SEISMICITY

THE following tables are of such value that it has been thought desirable to publish them entire. They constitute a great advance in the branch of seismology which treats of earthquake distribution. They are published in *Beiträge zur Geophysik*, vol. iv., under the title, "Introduction à un essai de description sismique du globe et mesure de la sismicité." The observations are brought down to the year 1897 in those districts in which the seismic occurrences have been most frequent and best observed and include in all 131,292 quakes and 10,499 epicentres. Especial care has been taken not to use the same quake twice and to exclude from each district all quakes which originate elsewhere. With the growth of observation many of the results may require modification. The tables formulate only the results of existing data reliably observed and recorded.

TABLE OF SEISMIC REGIONS CLASSIFIED BY GRAND DIVISIONS OF THE WORLD AND NUMERICAL DATA RELATING TO THEIR SEISMICITY

DESIGNATION OF COLUMNS

1. Number assigned to the district in each grand division.
2. Names of the districts.
3. Number of epicentres or of localities so considered.
4. Number of quakes reported from the district.
5. Period covered by the observations.
6. Mean annual frequency { h—historic. l—seismologic. g—seismographic.
7. Seismologic seismicity { computed from historic frequency sh. observed........................ sl. computed from seismogr. frequency sg.

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE I.—TERRES ARCTIQUES.									
1	Groenland......	4	6							
2	Islande.........	31	140	1839–90	5.56			67		
3	Ile Jean Mayen..	1	3							
	Totaux.....	36	149		5.56					
	EUROPE.									
	CHAPITRE II.—SCANDINAVIE.									
1	Suede septentrionale.........	42	116	1889–95		3.50			267.8	
2	Suede centrale....	65	109	1888–96		3.88			192.7	
3	Suede meridionale...........	16	25	1893–97		3.80			108.4	
4	Ile de Bornholm.	1	5							
	Suede-seismes generaux ou mal determines	5	5							
5	Christiania......	39	65	1886–95		2.50			177.4	
6	Cotes sud-occidentales de la Norvege......	56	116	1886–95		5.70			106.5	
7	Trondhjem......	14	19	1887–95		1.00			227.9	
8	Norrland (norvegien) et iles Loffoten	25	182	1819–29 1887–95		7.40			91.4	
9	Finmark et Lapland norvegiens.........	3	5	1887–95		0.57			327.2	
	Norvege-seismes generaux ou mal determines	4	9							
	Totaux.....	270	656			28.35				
	CHAPITRE III.—ILES BRITANNIQUES.									
1	Iles Shetlands...	2	5							
2	Ecosse du nord-est...........	5	11							
3	Canal Caledonien...........	27	54	1852–71	0.72			52		
4	Perthshire......	21	465	1852–90	0.64			35		
5	Basse Ecosse....	13	26	1886–89	1.25			31		
6	Angleterre septentrionale et centrale.......	68	167	1833–73	1.63			44		
7	Pays de Galles...	18	32	1839–94	0.28			68		
8	Cambridge......	10	18	1848–71	0.22			86		
9	Cotes anglais de la Manche.....	53	138	1848–71	1.91			49		
10	Irlande du sud-ouest.........	12	19	1852–80	0.31			65		
	Seismes generaux ou mal determines......	4	104							
	Totaux.......	233	1,139		6.96					

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE IV.—FRANCE (DU RHIN AUGOLFE DE GASCOGNE).									
1	Belgique........	20	63	1846–79	0.54			77		
2	Le nord.........	24	46	1857–74	0.27			167		
3	Cotes franc. de la Manche et iles normandes	38	90	1842–91	1.00			84		
4	Bretagne.......	23	43	1843–93	0.45			142		
5	Vendee.........	50	133	1877–94	1.55			66		
6	Le centre.......	35	57	1837–74	0.63			214		
7	Moselle moyenne Hardt et Luxembourg......	10	21	1866–84	0.22			189		
8	Lorraine et Pfalz.	9	10							
9	Alsace..........	24	91	1835–59 1875–93 1895–97	0.79			50.5		
10	Franche-Comte..	38	134	1838–93	0.88			86		
11	Dauphine et Savoie........	64	407	1842–93	2.48			47		
12	Alpes et Provence.........	20	71	1857–89	1.00			73		
13	Alpes Maritimes.	17	1,032	1859–61 1866–68 1870–71			138.58			15
14	Drome, Vivarais, et Vaucluse....	37	173	1835–89	0.91			51		
15	Chaine des Puys d'Auvergne .	21	70	1833–89	0.04			65		
16	Les Cevennes...	35	59	1837–94	0.52			188		
17	Hautes et basses Pyrénées.......	54	222	1849–85	4.79			26		
18	Le Sud-ouest ...	23	35	1847–75	0.48			215		
	Seismes generaux ou mal determines......	9	36							
	Totaux.....	551	2,793		16.55		138.58			
	CHAPITRE V.—LE PENINSULE IBERIQUE.									
1	Galice et Portugal...........	34	62	1841–89	0.76			173		
2	Navarre et Pays basques.......	16	17	1885–91	1.43			50		
3	Catalogne.......	23	44	1845–62 1883–87	1.35			66		
4	Espagne centrale	27	77	1841–61 1885–92	0.90			97		
5	Embouchure du Tage	13	203	1758–00 1838–91	0.89			55		
6	Andalousie interieure.......	19	97	1834–88	0.97			102		
7	Malaga.........	23	658	1834–88	2.46			31		
8	Almeria........	13	441	1851–65 1882–93	1.53			33		
9	Valence et Murcie........	29	1,057	1857–65		12.77			38	
	Totaux.....	197	2,656		10.29	12.77				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE VI.—SUISSE.									
1	Le Jura suisse...	47	338	1650–53, 1876–97		4.65			30.0	
2	La plaine suisse.	85	407	1876–97		7.09			34.4	
3	Les lacs suisses..	104	952	1879–97		6.26			43.1	
4	Cote nord du lac Leman ou de Geneve	27	217	1876–97		6.45			12.0	
5	Bas-Valais......	24	115	1879–97		3.58			25.3	
6	Haut-Valais	32	1,401	1856–71, 1880–97		3.14			13.9	
7	Les Grisons.. ..	35	119	1879–97		4.78			36.7	
8	Engadine.......	37	160	1879–97		5.05			21.6	
	Seismes generaux ou mal determines......	10	186							
	Totaux.....	401	3,895			41.00				
	CHAPITRE VII.—DU RHIN A LA VISTULE.									
1	Hollande.......	15	31	1824–54	0.48			84		
2	Jutland.........	16	25	1841–89	0.38			83		
3	Cotes de la mer du nord et de la Baltique de Dortmund a Dantzig.......	33	48							
4	Westphalie......	29	139	1846–83	1.49			40		
5	Taunus et Hundsruck, ou Nassau...........	32	107	1841–90	1.56			34		
6	Thuringe.......	10	18	1827–87	0.18			128		
7	Harz...........	18	40	1823–85	0.54			74		
8	Erzgebirge et Fichtelgebirge.	80	605	1850–84	1.43			47		
9	Riesengebirge...	7	13	1878–83	1.00			24		
10	Silesie..........	8	22	1875–78	1.25			42		
11	Bade...........	43	162	1888–97		2.00			66.9	
12	Odenwald.......	43	700	1875–83	1.79			26		
13	Wurtemberg et Souabe.......	68	189	1867–95		2.44			114.4	
14	Augsbourg......	5	27	1756–78, 1819–42	0.13			161		
15	Baviere orientale	11	15	1852–69	0.28			218		
16	Boheme.........	12	31	1854–71	1.85			78		
17	Plankerswald....	10	77	1874–77	1.50			30		
	Seismes generaux ou mal determines......	21	77							
	Totaux.....	457	2,326		13.86	4.44				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE VIII.—ALPES ORIENTALES.									
1	Vintschgau......	8	29	1874–79 1896–98		2.11			29.1	
2	Vorarlberg......	16	35	1897–98	5.50			5.7		
3	Tirol...........	36	185	1896–98		14.66			26.7	
4	Salzbourg.......	16	72	1897–99		6.00			42.4	
5	Autriche propement dite.....	45	128	1896–98		7.66			17.8	
6	Trentin.........	24	97	1873–91 1896–98		3.68			43.9	
7	Pusterthal......	11	19	1896–98		1.66			14.3	
8	Carinthie.......	47	190	1896–98		9.00			35.8	
9	Murthal.........	22	73	1896–99		13.00			21.9	
10	Murzthal et Soemering........	16	40	1898–99		2.00			26.2	
11	Styrie..........	35	68	1896–98		7.00			33.4	
12	Carniole........	134	495	1895–98		120.00			10.0	
13	Goritz..........	36	157	1896–98		44.33			9.1	
14	Istrie..........	29	276	1897–98		12.50			31.8	
	Seismes generaux ou mal determines......	17	31							
	Totaux.....	492	1,995			249.10				
	CHAPITRE IX.—CARPATHES ET MOYEN DANUBE.									
1	Raab...........	10	22	1838–76	0.41			123		
2	Sumegie........	10	43	1874–83	1.80			59		
3	Bakony-wald....	24	234	1849–70	2.77			32		
4	Moravie........	6	9	1858–65	0.37			134		
5	Trencsin........	8	173	1846–74	0.13			72		
6	Sohl...........	5	20	1854–69	0.87			21		
7	Abaujvar.......	20	51	1851–70	1.09			71		
8	Marmaros.......	8	26	1856–60 1867–83	1.14			17		
9	Altland.........	10	44	1851–86	0.81			56		
10	Jazygie.........	14	52	1863–72	5.30			18		
11	Torontal........	16	45	1852–87	0.97			52		
12	Banat..........	14	95	1858–71 1879–83	0.92			38		
13	Esclavonie......	9	26	1854–67 1876–84	0.98			51		
14	Croatie.........	15	240	1851–80	1.00			38		
	Seismes generaux ou mal determines......	15	28							
	Totaux.....	184	1,108		18.55					
	CHAPITRE X.—ITALIE CONTINENTALE.									
1	Alpes Cottiennes.	49	539	1873–93		7.95			26	
2	Piemont........	17	298	1873–93		7.78			19	
3	Doria Baltea....	20	72	1873–86		3.43			42	
4	Tessin italien....	25	109	1887–93		3.57			36	
5	Tessin suisse....	19	28	1877–82 1891–97		0.85			62.4	

1	2	3	4	5	6			7		
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6	Milanais........	19	67	1873–93		0.95			93	
7	Du lac de Come au lac de Garde	41	121	1887–93		6.14			42	
8	Verone, sette et trece comuni..	75	1,386	1873–86			61.00			18
9	Bellune ou Alpes dolomitiques...	35	376	1873–86			15.80			27
10	Frioul..........	24	226	1887–93		7.71			24	
11	Venetie.........	23	464	1873–93			15.14			63
12	Po inferieur.....	45	356	1873–93			8.38			12
13	Bolonais........	64	2,387	1873–87			82.35			23
14	Parmesan.......	40	358	1887–93			6.43			43
15	De la Serivia au Taro	20	189	1832–39 1840–55	4.09			19		
16	Du Tanaro a la Serivia........	32	200	1887–93			5.85			71
17	Riviere du Ponant..........	18	383	1888–93			11.33			33
18	Riviere du Levant..........	18	399	1887–89			58.33			32
	Totaux.....	584	7,958		4.09	38.38	264.61			
	CHAPITRE XI.—ITALIE PENINSULAIRE.									
1	Alpes Apuanes..	23	180	1873–86		0.64			68	
2	Toscane........	95	1,485	1873–93		21.19			18	
3	Archipel Toscan.	1	1							
4	Littoral toscan..	20	130	1875–81 1886–91		3.07			55	
5	Les lacs latins...	22	186	1887–93			3.50			42
6	Ombrie.........	77	2,407	1873–86			100.12			16
7	Les Marches.....	84	591	1873–87			21.40			35
8	Abruzze centrale ou Aquila.....	20	377	1884–93			21.30			47
9	Sabine..........	7	31	1884–97		1.14			53.3	
10	Albanese........	30	6,710	1873–87			388.87			4
11	Latium.........	59	777	1873–86			28.35			27
12	Molise et Campobasso.......	38	102	1873–91		2.57			54	
13	Capitanate et Monte Gargano	45	980	1885–92			12.16			33
14	Benevent.......	29	149	1883–86			13.00			45
15	Vesuve et Campanie.........	29	1,349	1859–87			35.62			19
16	Archipel Napolitain..........	9	198	1880–86			12.28			23
17	Littoral Napolitain ou Lucanie.........	15	55	1846–57	1.33			42		
18	Basilicate.......	34	311	1873–93		2.71			35	
19	Terres de Bari et d'Otrante.....	38	198	1884–93			11.90			96
20	Calabre septentrionale.......	27	148	1879–93		2.46			53	
21	Calabre centrale.	46	1,006	1850–76			20.51			39
22	Calabre meridionale.........	52	2,230	1843–53			26.63			39
	Calabre-seismes generaux ou mal determines	2	13							
	Totaux.....	802	19,614		1.33	33.78	695.55			

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE XII.—SICILE ET ILES ADJACENTES.									
1	Iles Eoliennes ou Lipari........	6	116	1887–93			10.57			26
2	Messine ou Monts Peloritani.....	25	731	1888–93			13.33			34
3	Etna...........	43	1,462	1873–87			55.42			4
4	Cote de Syracuse	27	587	1878–93			28.12			44
5	Cote de Girgenti	21	886	1876–87			55.58			33
6	Cote de Palerme.	22	181	1817–31, 1843–52, 1873–93	1.97			24		
7	Iles du canal d'Afrique (Julia, Malte, Pantellaria, etc.)..	18	154							
	Italie - seismes generaux ou mal determines	37	214							
	Totaux.....	199	4,331		1.97		107.60			
	CHAPITRE XIII.—BASSIN OCCIDENTAL DE LA MEDITERRANEE									
1	Baleares........	5	37							
2	Corse...........	4	5	1827–80	0.04			230		
3	Sardaigne.......	4	8	1835–76	0.14			205		
	Totaux.....	13	50		0.18					
	CHAPITRE XIV.—BALKANS ET BAS-DANUBE.									
1	Dalmatie.......	61	947	1843–60, 1896–98		26.80			29.8	
2	Bosnie..........	21	72	1872–88	3.92			55		
3	Herzegovine et Montenegro (Tschernagora)	8	37	1872–88	1.64			57		
4	Albanie, Epire et Corfou	46	893	1855–62, 1864–75, 1894–97		30.12			36.6	
5	Macedoine......	30	257	1855–61, 1864–76, 1894–97		7.37			92.6	
6	Roumelie Turque ou Thrace Bassin de la Maritza	19	154	1855–76, 1894–97		1.34			223.7	
7	Bulgarie........	14	60	1858–97		0.90			294	
8	Serbie..........	8	17	1889–95		1.57			198.9	
9	Valachie........	15	65	1855–97		0.77			324.5	
10	Moldavie et Bessarabie........	25	114	1854–95		1.43			224.8	
11	Galicie et Bukowine..........	12	16	1871–81	0.63			190		
	Totaux.....	259	2,632		6.19	70.30				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE XV.—GRECE.									
1	Thessalie.......	13	76	1863, 1867–68, 1895–97		7.00			42	
2	Eubee et Sporades du nord	23	1,228	(Eubee) 1857–78, 1895–97		44.92			9.6	
3	Attique, Parnasse, et Locride	43	1,979	1858–78, 1895–97		29.27			22.1	
4	Acarnanie..... .	17	138	1895–97		25.26			21.	
5	Iles Ioniennes...	41	5,700	1825–68, 1875, 1892–93, 1895–97		89.21			6.5	
6	Achaie.........	22	308	1860–76, 1882–83, 1887–88, 1895–97		15.83			20.8	
7	Corinthie et Argolide	28	311	1858–78, 1886–88, 1895–97		27.00			16.5	
8	Laconie	12	54	1858–62, 1867, 1876–77, 1895–97		3.16			45.8	
9	Messenie........	21	93	1895–97		3.33			30.7	
10	Arcadie.........	20	75	1895–97		2.66			33.7	
11	Crete ou Candie.	8	100	1858–88	1.64		70.7			
12	Les Cyclades....	14	141	1860–63, 1867–74, 1895–97		5.78			60.0	
	Seismes generaux ou mal determines.........	9	32							
	Orient - seismes generaux ou mal determines	9	71							
	Totaux.....	280	10,306		1.64	253.42				
	CHAPITRE XVI.—RUSSIE.									
1	Laponie russe...	8	22	1750–72, 1811–82	0.20					
2	Finlande........	32	92	1750–92, 1800–05, 1823–26, 1843–59, 1877–84	0.80			205.2		
3	Archipel d'Aland et golfe de Bothnie.......	3	5	1823–82	0.09					
4	Russie d'Europe proprement dite	56	104	1819–27, 1867–75, 1881–88	1.65					
5	Oural..........	12	30	1788–1888	0.21			276.5		
	Seismes generaux ou mal determines......	5	5							
	Totaux.....	116	258		2.95					
	Europe Totaux...	5008	61,717		90.18	731.54	1206.34			

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	ASIE.									
	CHAPITRE XVII. —SIBERIE.									
1	Basse Siberie....	9	14							
2	Altai...........	35	62	1761–66 1783–87 1822–31 1844–51 1879–87	1.18			322.2		
3	Baikalie........	18	248	1856–76	3.19			145.6		
4	Transbaikalie....	15	419	1847–87	9.54			72.8		
5	Territoires de l'Amour	12	42	1860–68 1883–88	1.47					
6	Ile Saghalien....	2	2							
7	Cotes de la mer d'Ochotsk.....	6	34							
	Totaux.....	97	821		15.38					
	CHAPITRE XVIII. —TURKESTAN.									
1	Caspienne orientale..........	7	12	1876–95	0.50					
2	Syr-Daria.......	12	348	1866–88	3.74			116.1		
3	Issyk-Koul......	21	297	1881–89	32.99			37.8		
	Totaux.....	40	657		37.23					
	CHAPITRE XIX. —ISTHME CAUCASIQUE.									
1	Cotes orientales de la mer noire.	12	42	1869–88	1.60			51.6		
2	Kouban........	14	42	1865–85	1.67			57.5		
3	Terek..........	31	127	1822–86	1.21			70.8		
4	Daghestan......	13	53	1841–55 1863–85	1.22			64.4		
5	Kour-Rive gauche	27	324	1801–05 1840–88	5.17			31.4		
6	Kour-Rive droite	20	107	1853–88	2.27			51.6		
7	Araxe-Rive gauche	23	131	1840–56 1868–74 1888–92	3.07			45.7		
8	Araxe-Rive droite.........	11	125	1840–83	2.29			57.1		
	Seismes generaux ou mal determines.........	13	23							
	Totaux.....	164	974		18.50					
	CHAPITRE XX.—ASIE MINEURE.									
1	Armenie turque..	29	247	1895–97		26.33			51.3	
2	Kyzil-Ermak....	33	132	1895–97		8.12			95.8	
3	Mer de Marmara et iles d'Imbros, Lemnos et Metelin	83	2,097	1855–78 1894–97		18.78			60.6	

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
4	Meandre ou Aidin et iles de Samos a Rhodes	79	1,801	1855–78 1883–88 1894–97		26.20			43.3	
5	Taurus.........	19	64	1895–97		4.00			140	
6	Ile de Chypre...	8	87							
	Seismes generaux ou mal determines.........	1	23							
	Totaux.....	252	4,451			83.43				
	CHAPITRE XXI. —L'ASIE DESERTIQUE.									
1	Syrie et Palestine..........	28	195							
2	Arabie et mer rouge orient...	8	19							
3	Mesopotamie....	13	48							
4	Farsistan, Mekran, et Beloutschistan meridional.........	12	66							
5	Iran...........	12	17							
6	Ghilan et Mazenderan.........	6	50							
7	Khorassan......	8	21							
	Totaux.....	87	417							
	CHAPITRE XXII. —INDE.									
1	Pendjab et Afghanistan.....	16	68							
2	Haut Pendjab et Cachemire.....	7	45	1885–86	17.50			20		
3	Kemaon, Nepaul, et Sikkim	13	37	1828–33 1842–43	2.25			104		
4	Assam..........	21	433	1874–80		26.00			34	
5	Gange et Bengale	17	49	1870–72	4.00			115		
6	Kutch et Sindh..	11	86	1841–56 1864–70	0.48			96		
7	Gudzerate et Bombay......	17	53	1868–72	5.60			66		
8	Deccan.........	11	15	1865–73	0.88			251		
9	Ceylon.........	8	12							
10	Golfe de Bengale.	3	3							
11	Iles Maldives et Tschagos......	1	1							
12	Seismes generaux ou mal determines.........	5	11							
	Totaux.....	130	813		30.71	26.00				
	CHAPITRE XXIII.— L'INDO-CHINE.									
1	Arracan et Birmanie........	7	19							

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
2	Presqu'ile de Malacca.........	3	11							
3	Indo-chine (Tonkin, Annam, Cochinchine, Cambodge, Siam).........	6	7							
4	Iles Andaman, Nicobar et Barren........	5	124							
	Totaux.....	21	161							
	CHAPITRE XXIV.—LE PLATEAU CENTRAL ASIATIQUE.									
1	Mongolie et Mandchouri...	4	8							
2	Dzoungarie chinoise, Turkestan chinois et Pamir........	8	89							
3	Kan-Sou........	10	112							
4	Thibet..........	4	5							
5	Yun-nan........	19	129							
	Totaux.....	45	343							
	CHAPITRE XXV.—CHINE.									
1	Leao-Toung et Coree.........	12	50							
2	Chine septentrionale.........	112	860							
3	Chine occidentale..........	43	853							
4	Chine centrale...	23	245							
5	Chine meridionale...........	24	66							
6	Ile d'Hainan....	1	14							
7	Mer de chine meridionale....	2	2							
	Seismes generaux ou mal determines.........	14	467							
	Totaux.....	231	2,557							
	CHAPITRE XXVI.—JAPON.									
1	Les Kouriles....	12	48	(1885–92) (Les Kouriles meridionales)	2.50				58.8	
2	Nemuro........	30	443			43.00			16.3	
3	Reste de l'ile d'Yeso........	3	344			5.12			111.4	
4	Siribesi.........	20	53			5.25			38.6	
5	Detroit de Tsugaru..........	47	301			23.75			24.3	

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
6	Nord-ouest du Nippon.......	50	131			9.37			58.3	
7	Rikiutsiu.......	43	475			15.12			36.0	
8	Golfe de Sendai..	58	286			17.87			28.3	
9	Bassin du haut Akanagawa....	18	71			2.75			42.2	
10	Bassin du Sinan-ogawa........	75	435			30.50			26.2	
11	Golfe du Toya-ma-Wan......	13	21			2.12			81.6	
12	Ouest de la presqu'ile de Noto.	30	128			15.37			25.7	
13	Pentes nord des plaines de Tokyo........	52	178			19.74			22.2	
14	Collines du Tsu-kuba-San.....	51	611			65.62			10.5	
15	Presqu'ile d'Awa et Kasuza.....	36	263			29.75			17.6	
16	Plaines de Tokyo	72	2,418			92.25			12.0	
17	Pentes ouest des plaines de Tokyo........	19	56			4.50			35.2	
18	Bassin du Fugi-gawa.........	33	92			7.75			29.1	
19	Bassin du Ten-riguawa.......	26	49			5.50			44.4	
20	Bassin du haut Kisogawa.....	36	417			14.37			21.3	
21	Mino...........	17	129			11.50			13.2	
22	Owari..........	48	3,356			24.74			18.5	
23	Ize............	24	95			9.37			24.0	
24	Sud de la presqu'ile de Kii...	39	212			25.00			28.4	
25	Lac Biwa et mer d'Izumi.......	67	1,934			16.75			26.0	
26	Golfe de Wakasa.	25	55			6.25			30.1	
27	Inaba, Tajima et Hoki oriental..	6	10			0.75			60.9	
28	Versant nord des mers d'Harima et de Bingo....	30	91			5.62			41.8	
29	Mers d'Harima et de Bingo......	17	43			5.12			28.9	
30	Sikoku sud-oriental.........	18	953			4.00			60.6	
31	Sikoku nord-oriental.......	7	9			1.00			54.4	
32	Versant coreen du Nippon sud-occidental.....	74	175			20.87			34.0	
33	Versant nord des mers de Suo, Iyo et Misima.	42	82			6.87			31.4	
34	Mers de Suo, Iyo et Misima.....	30	151			16.50			25.8	
35	Sikoku nord-occidental.....	10	24			2.87			33.7	
36	Sikoku sud-occidental........	15	36			3.87			41.6	
37	Versant oriental de Kiushiu....	32	75			7.87			34.4	
38	Kagoshima......	25	486			22.37			21.3	

1	2	3	4	5	6 h	6 l	6 g	7 sh	7 sl	7 sg
39	Kiushiu occidental..........	58	1,206			11.87			47.5	
40	Kiushiu nord-oriental.......	11	32			3.50			61.8	
41	Archipel Liou-Kiou (Tsubu-soto).........	4	35							
42	Formose (Thay-Ouan)........	12	49							
43	Archipels Bonin (Ogasawara Sima) et de L'archeveque..	3	4							
44	Iles Bayonnaise et Smith......	2	2							
	Seismes generaux ou mal determines.........	13	304							
	Totaux.....	1,359	16,368			605.83				
	Asie: Totaux .	2,426	27,562		101.82	631.83				
	AFRIQUE.									
	CHAPITRE XXVII.—ETATS BARBARESQUES.									
1	Maroc..........	12	26							
2	Tlemcen.........	3	6							
3	Oran...........	16	93	1886–94	3.11			81		
4	Tenes..........	3	43	1881–88	1.87			33		
5	Vallee du Cheliff.	10	31	1882–90	1.44			90		
6	La Mitidja......	12	215	1881–94	3.14			41		
7	Kabylie.........	13	48	1885–91	4.00			48		
8	Aumale et Dihra.	16	203	1885–88	20.50			31		
9	Constantine.....	18	135	1883–88	3.50			89		
10	Aures et Tell....	8	30	1885–94	0.50			184		
11	Tunisie.........	16	47							
12	Tripolitaine et Fezzan.........	3	4							
	Algerie. Seismes generaux ou mal determines	5	34							
	Totaux.....	135	915		38.06					
	CHAPITRE XXVIII.—AFRIQUE PROPREMENT DITE.									
1	Egypte et Ethiope...........	12	59							
2	Abyssinie et Erythree........	13	37							
3	Haut Nil et les grands lacs....	10	20							
4	Zanzibar........	1	1							

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
5	Nyassa et du Zambeze a la Limpopo......	4	7							
6	Afrique meridionale, du Cap a la Limpopo..	8	21							
7	Cotes de l'Orange au Congo	1	1							
8	Congo..........	1	1							
9	Golfe de Guinee (de l'embouchure du Congo au cap Palmas).	2	12							
10	Senegal.........	2	5							
11	Sahara.........	2	15							
	Totaux.....	56	179							
	CHAPITRE XXIX.—OCEAN ATLANTIQUE.									
1	Les Acores......	12	1,444	1862–67	6.22			20		
2	Madere.........	1	10							
3	Les Canaries....	13	67							
4	Iles du cap vert.	3	11							
5	Rocher St. Paul.	33	35							
6	Iles Ascension, Sainte Helene, Tristan d'Acunha et Atlantic sud..........	13	29							
7	Atlantique nord.	3	3							
8	Atlantide.......	33	42							
9	Atlantique (des iles du cap vert a l'Amazone)..	12	12							
10	Region volcanique de Daussy (Atlantique equatorial)....	39	51	1831–51	25.00			201		
	Totaux.....	162	1,704		31.22					
	CHAPITRE XXX.—OCEAN INDIEN.									
1	Les Mascareignes	8	16							
2	Madagascar.....	6	25							
3	Les Comores....	1	2							
4	Iles Kerguelen, St. Paul et Amsterdam (pour memoire).....									
5	Ocean indien....	14	14							
	Totaux.....	29	57							
	Afrique: Totaux	382	2,855		69.28					

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	AMERIQUE DU NORD.									
	CHAPITRE XXXI.—MER DE BEHRING.									
1	Kamtchatka et iles du Commandeur......	12	98	1841–54	4.00			87.5		
2	Aleoutes........	15	86							
3	Alaska.........	7	12							
	Totaux.....	34	196		4.00					
	CHAPITRE XXXII.—CANADA OU DOMINION.									
1	Labrador et Terre-Neuve...	3	3							
2	Nouvelle Ecosse, nouveau Brunswick et ile du Cap Breton	13	13	1847–53 1884–86	1.00			137		
3	Saint Laurent...	23	80	1879–88	4.10			59		
4	Cotes nord des lacs Erie et Ontario.......	14	16	1877–85	1.22			80		
	Totaux.....	53	112		6.32					
	CHAPITRE XXXIII.—VERSANT ATLANTIQUE DES ETATS-UNIS.									
1	Nouvelle Angleterre	105	478	1727–41 1791–94 1876–85	16.47			90		
2	Les Carolines....	56	160	1849–61 1874–86	3.61			154		
3	Cotes sud des Lacs Erie et Ontario	32	55	1844–60 1865–73 1877–85	1.54			170		
4	Ohio, Tenn. et moyen Missip.	67	162	1846–56 1875–85	3.95			170		
5	Cotes du golfe du Mexique......	15	17							
6	Michigan.......	9	9	1847–55	2.20			240		
7	Le Far-West....	23	29							
	Etats-Unis et Canada. Seismes generaux ou mal determines.........	4	27							
	Totaux.....	311	937		27.77					

1	2	3	4	5	6		
					h	l	g
	CHAPITRE XXXIV.—VERSANT PACIFIQUE DES ETATS-UNIS.						
1	Cotes du mont St. Elie aux iles Scott......	7	97				
2	Washington et Vancouver	32	94	1877–85 1888–96		4.22	
3	Californie septentrionale.......	51	213	1877–96		4.90	
4	Calif. centrale...	113	1,096	1860–97		21.86	
5	Californie meridionale	62	747	1848–57 1878–96		9.55	
6	Les Montagnes Rocheuses.....	60	2,179				
	Calif. Seismes gener. ou mal determines....	19	41				
	Totaux.....	344	4,467			40.53	
	CHAPITRE XXXV.—MEXIQUE.						
1	Basse ou Vieille Californie			1845–57 1870–71 1878–92		1.61	
2	Sonora et Sinaloa..........	10	109	1887–93		15.57	
3	Desert de Mapimi...........	15	34	1880–87		1.82	
4	Tamaulipas.....	9	10	1880–87		0.44	
5	San-Luis Potosi et Queretaro..	18	75	1888–99, 5		3.33	
6	Ceboruco.......	6	36	1875–88		2.28	
7	Colima.........	20	189	1885–99, 5		18.88	
8	Chapala........	30	485	1872–93		17.59	
9	Flanc sud de la sierra Madre ou Acapulco.	27	471	1878–89 1895–99, 5		15.52	
10	Rio Mexcala....	48	457	1878–89 1895–99, 5		21.75	
11	Plateau central d'Anahuac ou de Mexico	18	339	1644–98 1784–05 1844–99, 5		15.70	
12	Oaxaca et Tehuantepec.......	29	1,134	1875–99, 5		42.12	
13	Orizaba.........	23	2,102	1845–49 1878–84 1887–90 1895–99, 5		5.61	
	Orizaba (observ. sismographiques)......			1887–95			217.77
14	Coatzacoalcos...	4	8	1871–90		0.65	
15	Chiapas et Tabasco	8	26	1872–87 1897–99, 5		1.13	

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
16	Tonala.........	3	5	1897–99, 5		1.20			84.9	
17	Archipels Revillagigedo et Tres Marias...	4	5							
	Seismes generaux ou mal determines.........	8	56							
	Totaux.....	288	5,586			159.59	217.77			
	CHAPITRE XXXVI.—LE CENTRE AMERIQUE.									
1	Belize et Honduras.........	15	53	1846–56	2.09			42		
2	Guatemala......	30	843	1853–63		12.91			21	
3	Salvador........	26	1,181	1881–84		44.50			13	
4	Nicaragua......	12	70							
5	Costarica.......	13	549	1866–80		18.00			17	
6	Darien.........	9	43	1882–88		4.14			28	
	Totaux.....	105	2,739		2.09	79.55				
	CHAPITRE XXXVII.—LES ANTILLES.									
1	Les Bermudes...	1	9							
2	Les Bahamas....	2	2							
3	Cuba occidental.	5	15	1862–63 1880–81	1.75			52		
4	Cuba central et les Caimans...	4	4	1851–57	0.28			218		
5	Cuba oriental...	11	192	1854–67	2.57			57		
6	La Jamaique....	17	157	1847–73	2.48			32		
7	Haiti et Saint Domingue.....	22	262	1783–89	5.14			59		
8	Portorico et iles Vierges	18	694	1864–66 1869–79	8.00 4.00			23 32		
9	Pet. Antilles ou iles du vent...	40	1,196	1845–71	7.03			42		
10	Mer interieure des Antilles....	2	2							
11	Fosse maritime a l'est des Antilles...........	12	15							
	Seismes generaux ou mal determines.........	2	13							
	Totaux.....	136	2.561		27.25					
	Amerique du nord: Totaux.	1,271	16,598		67.43	279.67	217.77			

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	AMERIQUE DU SUD.									
	CHAPITRE XXXVIII.—VERSANT ATLANTIQUE DE L'AMERIQUE DU SUD.									
1	Orenoque.......	10	26							
2	Les Guyanes....	8	25							
3	De l'Orenoque au cap Horn.....	19	28							
4	Tucuman.......	13	265	1873–86		7.64			50	
	Totaux.....	50	344			7.64				
	CHAPITRE XXXIX.—ANDES DU NORD.									
1	Paria et Trinidad	9	113	1863–68	3.16			33		
2	Venezuela proprement dit ou Caracas	17	259	1857–71 1885–87		12.44			34	
3	Iles sous le vent.	2	6							
4	Andes du Venezuela	20	106	1865–70 1885–87		8.00				73
5	Basse Magdalena et Maracaybo	11	34	1845–56 1865–69 1883–86	1.94				193	
6	Haute Magdalena..........	12	79	1866–72	2.71				80	
7	Cauca et Atrato.	3	5							
8	Andes de Quito..	30	772	1864–72 1879	5.50				40	
9	Guayaquil......	8	48	1866–72 1878–80	3.40				74	
10	Les Gallapagos (pour memoire)........									
	Totaux.....	112	1,422		16.71	20.44				
	CHAPITRE XL.—ANDES DU CENTRE.									
1	Cote peruvienne septentrionale de Payta a Casm	6	19	1855–75	0.62			138		
2	Cote peruvienne centrale du Callao a Ica	14	1,177	1709 1858–65 1867–71		19.53			32	
3	Cote peruvienne meridionale de Caraveli a Iquique	18	1,621	1810–46 1862–72		31.27			53	
4	Haut Perou et Bolivie........	23	67							
	Totaux.....	61	2,884		0.62	50.80				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE XLI.—ANDES DU SUD.									
1	Chili septentrional de Chanaral a Illapel	16	1,103	1849–73, 1875–86		36.51			20	
2	Chili central d'Illapel a Concepcion	31	1,512	1836–41, 1849–86		28.25			42	
3	Chili meridional de Concepcion a Puerto Montt	30	772	1867–72	3.83			42		
4	Extreme sud du Chili (de Puerto Montt au cap Horn).....	4	5							
5	Iles Juan Fernandez........	3	11							
	Seismes generaux ou mal determines.........	3	28							
	Totaux.....	87	3,431		3.83	64.76				
	Amer. du sud: Totaux.....	310	8,081		21.16	143.64				
	OCEANIE.									
	CHAPITRE XLII.—SUMATRA.									
1	Atchin ou Atjeh.	10	97	1884–91, 1895–96		7.45			37.5	
2	Tapanoelie et Poeloe-Nias	26	384	1852–68, 1870–96		7.93			28.3	
3	Padang.........	43	552	1845–46, 1850–68, 1870–96		9.73			46.9	
4	Bengkoelen.....	24	268	1854–67, 1876–98		7.27			57.4	
5	Ile Engano......	7	9	1853–93		0.17				
6	Palembang et Lampongs	28	165	1850–56, 1875–98		3.71			107.5	
7	Archipels orientaux (Bintang ou Riouw, Bangka, Billiton)..........	8	14							
	Seismes generaux ou mal determines.........	1	42							
	Totaux.....	156	1,559			36.26				
	CHAPITRE XLIII.—JAVA.									
1	Java occidental..	104	931	1846–49, 1851–63, 1870–98		10.98			43.9	
2	Java central occidental	79	535	1846–48, 1850–68, 1870–73, 1875–98		14.11				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
3	Java central oriental	89	510	1845–68, 1870–73, 1875–98		9.09			61.5	
4	Java oriental....	24	92	1884–98		3.40			59.2	
5	Madoera........	6	10	1887–95		0.55			88.8	
	Seismes generaux ou mal determines.........	5	41							
	Totaux.....	307	2,155			38.13				
	Chapitre XLIV.—Les Moluques.									
1	Borneo brittan..	1	2							
2	Borneo hollandais occidental.	5	9							
3	Borneo hollandais oriental...	5	10							
4	Celebes meridional (presqu'ile de Mangkasar) et iles Saleyer.	14	61	1885–98		2.93				
5	Celebes central et golfe de Tomaiki ou Tolo.	2	9	1896–98		2.66			109.8	
6	Celebes septen. (presqu'ile de Menado)......	35	1,069	1845–98		17.91			41	
7	Iles Sanguir, Siao et Talauer.....	7	27							
8	Moluques septen. (Morotai, Halmaheira, Ternate, Ombaai	16	981	1770–74, 1812–34, 1846–71, 1889–98		10.35			67	
9	Moluques merid. (Boeroe, Ceram, Saparoea, Amboine, Banda).	34	815	1841–98		12.55			41.6	
10	Iles de Wetter aux Aroe......	18	48	1886–98		5.20			218.1	
11	Timor..........	12	167	1856–60, 1864–68, 1893–97		5.20			76.1	
12	Iles de Bali a Allor.........	30	276	1859–98		7.60			98.4	
	Totaux.....	180	3,474			64.40				
	Chapitre XLV.—Les Philippines.									
1	Rio Grande ou Lucon nordest.	25	364	1870–97		4.22			86.1	
2	Les Ilocos.......	31	274	1867–97		7.26			43.9	
3	Manille.........	36	689	1869–97		8.25			18.8	
4	Iles Mindoro, Calamianes, Cuyos Palawan et Balabac	10	135	1871–78, 1890–97		3.13			53.8	

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
5	Presqu'ile des Camarines.....	16	493	1878–97		10.60			34.4	
6	Iles Burias, Masbate, Samar et Leyte.........	14	281	1869–97		2.31			114.3	
7	Iles Panay, Cebu et Bohol......	14	122	1869–97		2.79			128.1	
	Lucon. Seismes gen. ou mal det	5	41							
	Totaux.....	151	2,399			38.50				
	CHAPITRE XLVI.—MINDANAO.									
1	Mindanao nord-oriental.......	28	1,181	1878–98		21 80			27.9	
2	Mindanao sud-oriental	5	68	1870–73 1878–79 1891–98		4.76			66.8	
3	Mindanao cent..	6	131	1871–72 1888–98		5.50			87.9	
4	Presqu'ile de Zamboanga et ile Basilan.....	11	122	1869–98		2.79			73.9	
5	Archipel Soulou.	2	10							
6	Iles Palaos ou Pelew.........	1	1							
	Totaux.....	53	1,513			34.85				
	CHAPITRE XLVII.—AUSTRALIE ET TASMANIE.									
1	Australie.......	24	72							
2	Tasmanie.......	6	11	1859–84	0.45			192		
	Totaux.....	30	83		0.45					
	CHAPITRE XLVIII.—NOUVELLE ZELANDE.									
1	Auckland.......	20	125	1869–95		4.52			128.1	
2	Detroit de Cook.	30	1,540	1846–48 1868–95		14.97			72.2	
3	Canterbury.....	16	100	1868–95		3.14			137	
4	Otago et ile Stewart..........	9	55	1871–95		2.12			151.4	
5	Cotee Occi. de l'ite du milieu.	5	23	1870–95		0.88			145.3	
6	Iles Kermadec, Chatam et Auckland.....	3	3							
	Seismes gen. ou mal determ....	18	79	1868–95		2.39				
	Totaux.....	101	1,925			28.02				

1	2	3	4	5	6			7		
					h	l	g	sh	sl	sg
	CHAPITRE XLIX. —OCEANIE PROPREMENT DITE OU POLYNESIE.									
1	Iles Mariannes ou des Larrons...	6	246	1892–98		15.00				
2	Les Carolines....	1	2							
3	Nouvelle Guinee et nouvelle Bretagne......	9	34							
4	Nouvelles Hebrides, Santa Cruz et iles Salomon	12	52							
5	Nouvelle Caledonie et iles Loyalty	4	18	(Nouvelle Caledonie) 1875–87		0.61			161	
6	Iles Tonga, Samoa et Tokelau...........	11	62							
7	Iles Sandwich ou Hawai........	18	778	(Ile Hawai) 1843–74		13.0			30	
8	Polynesie et epicentres sporadiques de l'Ocean Pacifique.........	28	30							
	Totaux.....	89	1,222			28.61				
	Oceanie: Totaux.....	1,066	14,320		0.45	268.83				

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